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ARCHEAN LODE GOLD DEPOSITS IN ONTARIO

PART I. A DEPOSITIONAL MODEL

PART II. A GENETIC MODEL

by

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FOREWORD

The Ontario Geological Survey of the Ministry of Northern Development and Mines has, since 1980, mounted a substantial program to provide new, primarily field-based, descriptive data on the geological environment of Archean lode gold deposits. The early results of this program were reported in 1984 in Open File Report 5524 - An Integrated Model for the Origin of Archean Lode Gold Deposits, which was intended to promote discussion.

Subsequent work in this gold program has greatly increased the data base upon which the 1984 genetic model was developed. These new data have confirmed the principal premises of the early genetic model - that most, if not all, Archean gold deposits are fundamentally similar, and have resulted from a single set of genetic processes. The different characteristics of individual deposits result from the interaction of a simple set of fundamental variables upon a similarly simple set of inherent characteristics.

Part I of this paper presents this expanded base of descriptive data on Ontario's Archean lode gold deposits and an interpretation of that data. Part II of the report addresses the more theoretical problems of the origin of the gold and how it was transported to present sites of mineralization. In total, the data presented and the models derived from them offer significant new insight into how to explore for gold deposits in Ontario's Precambrian Shield.

V.G. Milne
Director
Ontario Geological Survey
2 May 1988



INTRODUCTION

In the four years since the publication of a preliminary version of this report (Colvine *et al.* 1984), extensive new descriptive data on Archean lode gold deposits have been generated; research into the genesis of gold deposits has also increased. In the past, the empirical, descriptive data on these deposits have proven of more practical use in exploration. For these data to be successfully applied in exploration, gold deposits must comply with consistent geological - metallogenetic controls; if they do not, then a geologically chaotic situation exists, in which data from any one deposit may be pertinent only to that deposit. Thus the practical and theoretical aspects are to a certain extent interdependent. An attempt is made in this report to deal principally with information pertinent to the depositional environment in Part I, and with more fundamental aspects of gold genesis in Part II.

Part I expands on the premise discussed in Colvine *et al.* (1984) that most, if not all, Archean lode gold deposits are fundamentally similar. Based on this premise, data from many individual deposits and mining camps are drawn together into a composite depositional model. The model attempts to explain the variability in many of the inherent characteristics of gold deposits by the interaction of a defined set of fundamental geological variables.

Inherent characteristics common to most gold deposits are their occurrence within regional scale linear composite shear systems, their apparently similar late Archean age over large areas of the Shield, and their spatial association with a late Archean suite of felsic intrusions. Form and distribution of gold mineralization and associated wallrock alteration, however, show a high degree of variability from one deposit to another.

The principal fundamental variables addressed are lithologies which host mineralization, and ambient metamorphic environment and hydrothermal fluid pressures during mineralization. On a local scale, variability in inherent characteristics can commonly be attributed principally to variability in host lithologies. On a larger scale, the variability in the inherent characteristics within a single gold system, or between systems, can be explained principally as a function of depth of burial beneath the paleosurface during mineralization. Metamorphic grade of the host rocks is an indicator of depth of burial, although pressure - temperature conditions may change during the life of a mineralizing system. This results in a vertical zonation of gold deposits, which is discussed extensively in Part I of the report. The effects of the mineralizing fluid itself may locally disrupt this overall systematic pattern.

Empirical evidence is presented throughout Part I for the consistent relative timing of mineralization and its absolute age is addressed directly. Similarly, evidence for the consistent composition of the fluid is presented indirectly from alteration studies and directly from fluid inclusion and isotope geochemistry data. The cause of gold precipitation to form mineable concentrations has not been definitely established, but several lines of evidence are discussed which help establish which processes may have been more important.

Colvine *et al.* (1984) presented evidence which was considered to be inconsistent with a model which invokes syngenetic concentration during volcanism and sedimentation, subsequently reconcentrated by structural, magmatic and metamorphic processes. This report does not address alternative models directly, but presents additional evidence that the observed characteristics of gold deposits can be explained as the product of a single mineralizing event, unrelated to volcanism and sedimentation. The test of this model is its internal consistency in demonstrating the interrelationships of the various geological processes which operated during mineralization.

Part II of the report discusses the more fundamental problems of where gold came from and how it was transported into its depositional sites. Evidence is presented in Part I that final transport was in hydrothermal fluid. The volumes of fluid indicated are too great to have been derived locally; the fluids and the gold must, therefore, have been derived from a source external to and deeper than the depositional sites. Part II, therefore, examines crustal scale processes which were temporally related to the gold mineralizing event.

Geochronological evidence is now adequate to demonstrate that the deformation zones which host gold deposits developed during a Shield-wide tectonic compression which occurred during a major magmatic event of batholith emplacement and a metamorphic event which affected the upper and lower crust. How these crustal processes may have interrelated to produce solution, transportation and deposition of gold is discussed, but it is clear that there is not yet a single definitive solution to this complex geological problem. What is clear, however, is that gold concentration is the product, or more correctly the byproduct, of a fundamental stage in the Earth's evolution which is manifested as final cratonization of the Archean crust.

ACKNOWLEDGEMENTS

Many people have helped in the preparation of this report. We thank the geologists of mining and exploration companies and the prospectors who have made us welcome on their properties, and have spent many hours showing us the geology of those properties. Many of our colleagues in the Ontario Geological Survey and the Regional Offices of the Ministry of Northern Development and Mines enthusiastically argued with us about the concepts presented in the report, helping to refine our ideas. Discussions with geologists from other government agencies and universities were equally important.

This report is the culmination of a program begun in 1980 to study Ontario's gold deposits. Contributions to that program have been made by many geologists other than the authors, and are contained in the report. We thank all of them.

We are indebted to many people in the Ontario Geological Survey, whose contributions of time and work allowed completion of the report in an extremely short period of time. In particular, critical reviews by Greg Stott, Richard Sutcliffe, Phil Thurston and Howard Williams of the Precambrian Geology Section made substantial improvements to the manuscript. Staff of the Precambrian Geology Section and of the Scientific Review Office of the Ontario Geological Survey were exceptionally helpful, devoting themselves to long hours of hard work to enable us to meet stringent publication deadlines.

Lastly, but most importantly, the authors thank Mike Cherry, who assumed editorial responsibility for this paper at short notice. His contributions are two-fold. As a geologist, he made significant contributions to the content of the paper. As editor, he assumed the onerous task of attending to the myriad scientific and technical details required to ensure the high quality of the finished product. Despite the pressures of an unyielding deadline and the demands of the authors, he maintained his approachability and good sense of humour. His contributions are gratefully acknowledged by all of the authors.

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PART I. A DEPOSITIONAL MODEL



1. THE GEOLOGICAL SETTING OF GOLD IN THE SUPERIOR PROVINCE

1.1. INTRODUCTION

A knowledge of the regional geological setting of Ontario's Archean iron ore gold deposits is prerequisite to any discussion of how these deposits may have formed. Deposit-scale lithological associations of gold provide critical empirical observations which are essential to construction of the descriptive model that is developed in Part I of this volume. This chapter, then, first presents summaries of the distribution of gold in the Superior Province, and of the geology of the Superior Province. These are followed by a summary of the lithological associations of gold, and the implications of those associations.

1.2. DISTRIBUTION OF GOLD DEPOSITS IN THE SUPERIOR PROVINCE

The Superior Province (Figure 1.1) is the largest exposed Archean craton in the world, and has accounted for more gold production than any other Archean craton, with more than 50 million ounces (4500 tonnes) produced to date. The locations of the principal gold camps in the Superior Province are shown on Figure 1.1. Table 1.1 lists the 25 largest deposits known to date: each of these has produced, or is known to contain, more than 1 million ounces (30 tonnes) of gold.

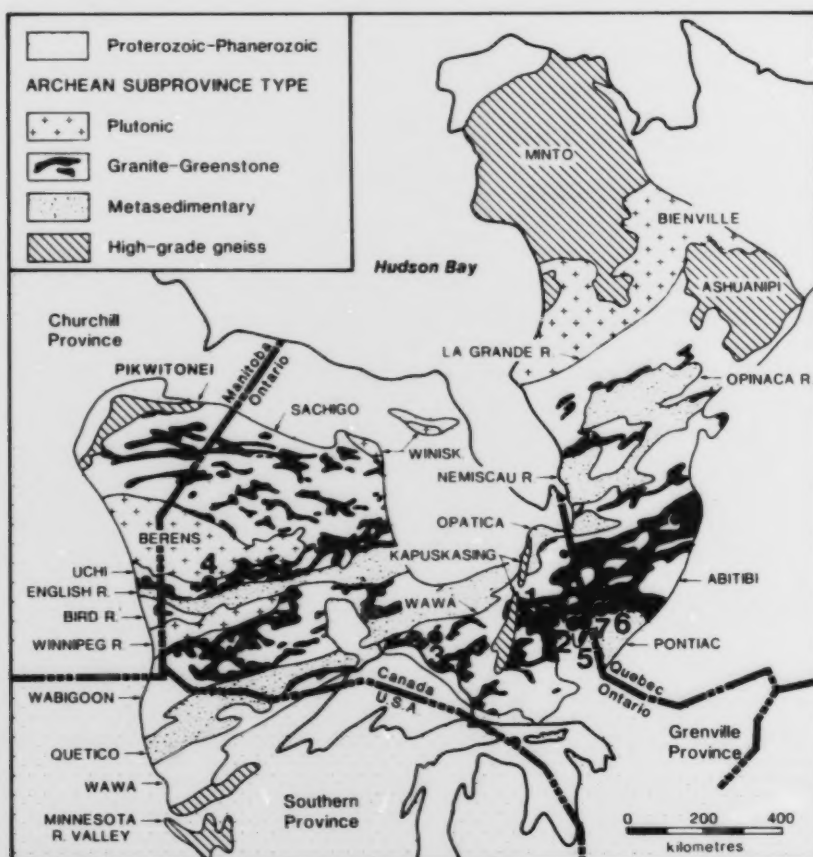


Figure 1.1. Generalized geology of the Superior Province, showing subprovinces and the locations of major gold camps. Gold camps shown include: 1 - Timmins/Porcupine, 2 - Kirkland Lake, 3 - Hemlo, 4 - Red Lake, 5 - Larder Lake, 6 - Val d'Or, and 7 - Malartic. Map modified from Card and Ciesielski (1986).

<u>MINE</u>	<u>GOLD</u> <u>(MILLIONS OF TROY OUNCES)</u>
1. Hollinger, Porcupine	19.4
2. Dome, Porcupine	10.8
3. McIntyre, Porcupine	10.6
4. Kerr Addison, Larder Lake	10.1
5. Page Williams, Hemlo	9.1
6. Lake Shore, Kirkland Lake	8.5
7. Golden Giant, Hemlo	6.7
8. Campbell Red Lake, Red Lake	5.5
9. Wright-Hargreaves, Kirkland Lake	4.8
10. Lamaque, Val d'Or	4.7
11. Teck-Hughes, Kirkland Lake	3.7
12. Sigma, Val d'Or	3.3
13. David Bell, Hemlo	3.0
14. Pamour (no.1), Porcupine	2.9
15. East Malartic, Val d'Or	2.9
16. Macassa, Kirkland Lake	2.6
17. Aunor (no.3), Porcupine	2.5
18. Madsen, Red Lake	2.4
19. Dickenson, Red Lake	2.3
20. Malartic Goldfields, Val d'Or	1.7
21. Sylvanite, Kirkland Lake	1.7
22. Hallnor (no.2), Porcupine	1.6
23. Preston, Porcupine	1.5
24. Camflo, Val d'Or	1.5
25. Detour Lake Mine, Detour Lake	1.5

Table 1.1. Gold content (production and/or reserves) of the 25 largest gold mines in the Superior Province.

As illustrated by Table 1.1, the Abitibi belt is clearly the most prolific gold-producing greenstone terrane in the Superior Province: the Timmins (Porcupine) camp alone has contributed in excess of 51 million ounces (1530 tonnes) of gold, far greater than production from the Kalgoorlie camp in Western Australia, or from the Homestake deposits in South Dakota. The recently discovered deposits at Hemlo contain proven reserves in excess of 18.8 million ounces: these reserves, which will undoubtedly increase as development continues, have already placed Hemlo within the top 5 gold camps of the Superior Province.

One prominent characteristic of all significant gold deposits in the Superior Province is their occurrence within or immediately adjacent to greenstone belts (Figure 1.1). They are not, however, preferentially hosted by a specific greenstone lithology or lithological assemblage, but occur within all greenstone lithologies. This is in marked contrast to the economically significant, volcanogenic massive sulphide deposits, which show a strong correlation with felsic pyroclastic rocks that are geochemically and geochronologically distinct.

A second, equally prominent, characteristic of the gold deposits is their occurrence within major tectonic zones which comprise linear composite shear systems (Figure 1.2). These shear systems, or deformation zones, are commonly of regional extent, exhibit systematic orientations and sense of shear, and may truncate all Archean lithologies. The deformation zones and their relationship to gold deposits are discussed in detail in the next chapter of this.

1.3. GENERAL GEOLOGY OF THE SUPERIOR PROVINCE

Card and Ciesielski (1986) divided the Superior Province into four major subprovince types (Figure 1.3): volcano-plutonic, plutonic, metasedimentary, and high grade gneiss. The boundaries of these subprovinces are either major dextral, transcurrent, east-striking faults (e.g. the Quetico Fault), or zones of structural and metamorphic transition (e.g. the boundary between the Winnipeg River and Wabigoon subprovinces).

The greenstone belts which host the gold deposits occur as east-northeasterly-trending, ribbon

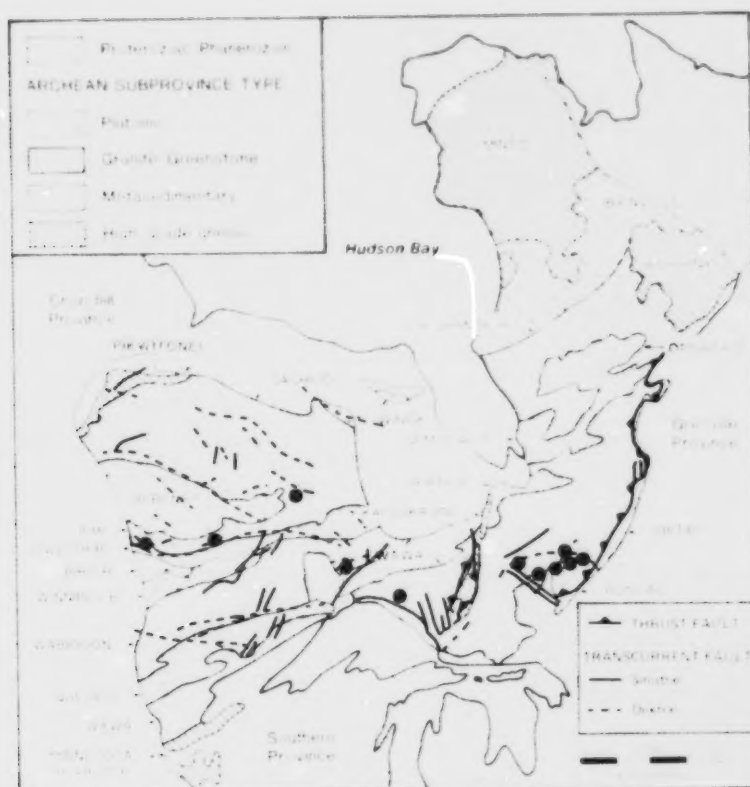


Figure 1.2. Major faults of the Superior Province, with major gold camps superimposed. See Figure 1.1 for explanation of gold camps. Map modified from Card and Ciesielski (1986).

or amoeboidal domains in the volcano-plutonic terranes (Figure 1.1). These belts have accounted for the majority of the wealth of metallic mineral production of the Superior Province. Wood and Wallace (1986) and Ayres *et al.* (1985) present comprehensive descriptions of the characteristics of the greenstone terranes of the Canadian Shield. The greenstone belts, which in general have been well preserved, typically consist predominantly of mafic to ultramafic and felsic metavolcanics, interlayered with clastic and chemical metasediments. In several of the belts, a volcanic cyclicity occurs on both minor and major scales (Thurston 1986). Locally, the volcano-sedimentary sequences are unconformably overlain by linear belts of shallow water to subaerial metasediments and alkalic volcanics, which have historically been called "Timiskaming-type" (Cook and Moorhouse 1969).

The supracrustal rocks have been intruded by syn-volcanic plutons, and are bordered by massive to foliated granitoid rocks and gneisses. In addition to these older intrusions, late Archean, saturated

and undersaturated felsic to mafic igneous rocks were intruded into the greenstone belts.

The metamorphic grade of the present erosional level of most greenstone terranes ranges from sub-greenschist to greenschist facies in the centre, to lower amphibolite facies at the margins, where the belts grade into gneissic terrains or are intruded by external granitoid batholiths. Amphibolite facies contact metamorphic aureoles occur around intrusions into the greenstones (cf. Jolly 1978, 1980).

The greenstone terranes have been subjected to internal folding and tilting, so that most strata are now subvertical. Strata most commonly face away from the marginal and internal granitoid bodies and form large synclinoria (eg. the Blake River Synclinorium in the Abitibi greenstone belt). Most workers have attributed this tilting to diapiric ascent of batholithic complexes (eg. Schwerdtner *et al.* 1979). However, strain aureoles are normally limited to discrete zones surrounding the batholiths, and large areas of greenstone terranes remain remarkably unaffected by the strain that can be

PART I. DEPOSITIONAL MODEL

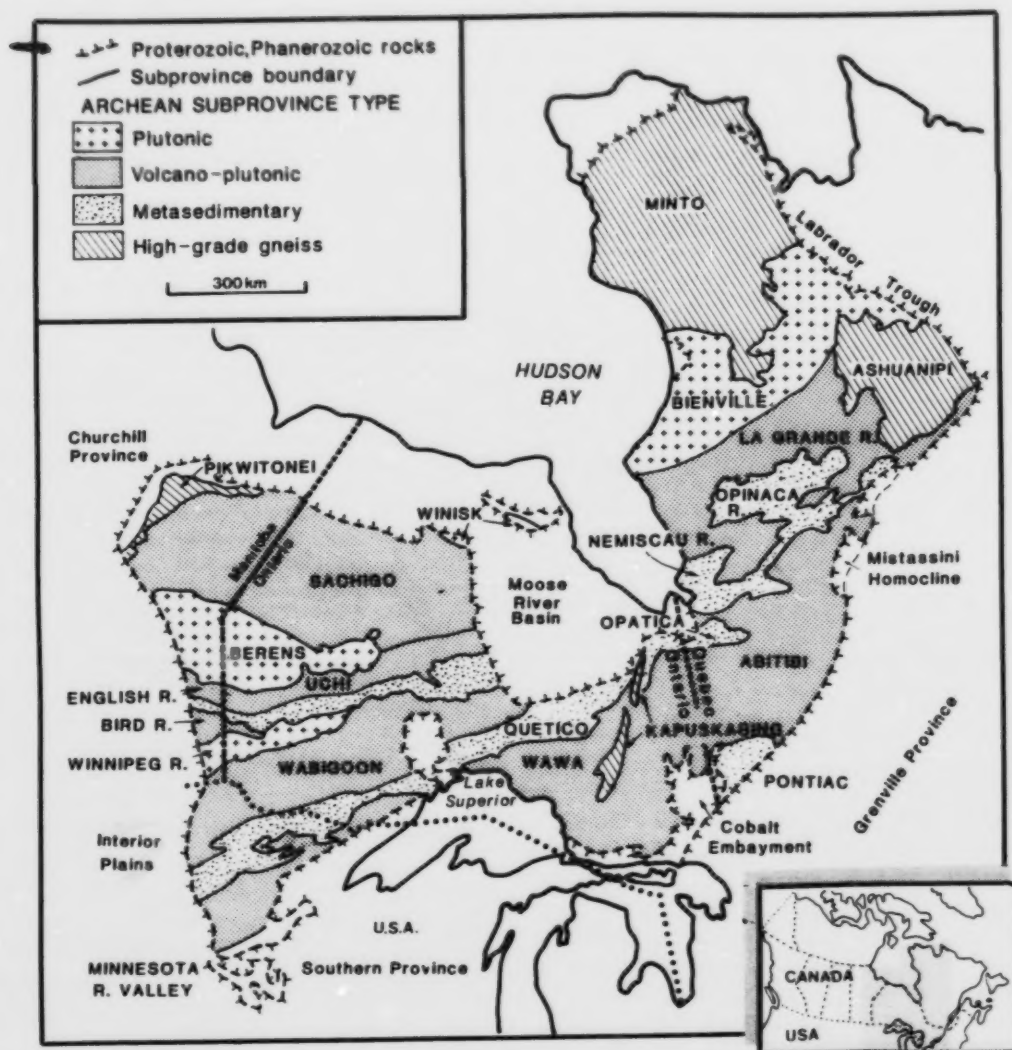


Figure 1.3. Distribution of major subprovinces in the Superior Province. Diagram after Card and Ciesielski (1986).

directly attributed to the emplacement of the large batholiths (Stott 1985). In the light of this evidence, and newly recognized stratigraphic repetitions and reversals in younging directions, other models, such as rifting and wrench faulting (Ludden *et al.* 1986), thrusting (Williams 1987, McGill and Shradly 1986), and/or listric normal faulting (Hodgson 1983b, Gibson *et al.* 1986), must be considered as possible mechanisms of tilting of the strata.

Structural and stratigraphic continuity are locally completely disrupted by late shearing, associated with the major deformation zones, as discussed in Chapter 2, Part I of this volume.

1.4. GEOCHRONOLOGY OF THE SUPERIOR PROVINCE

More high precision U-Pb zircon and titanite geochronology has been carried out on samples from the Superior Province than on any other Archean craton. These data, combined with high quality, detailed mapping, provide a comprehensive geochronological framework for the Superior Province. This absolute time framework is essential to understanding how the Superior Province developed. It is also an essential time frame with which the timing of gold mineralization, discussed in Chapter 5, Part I, must be compatible.

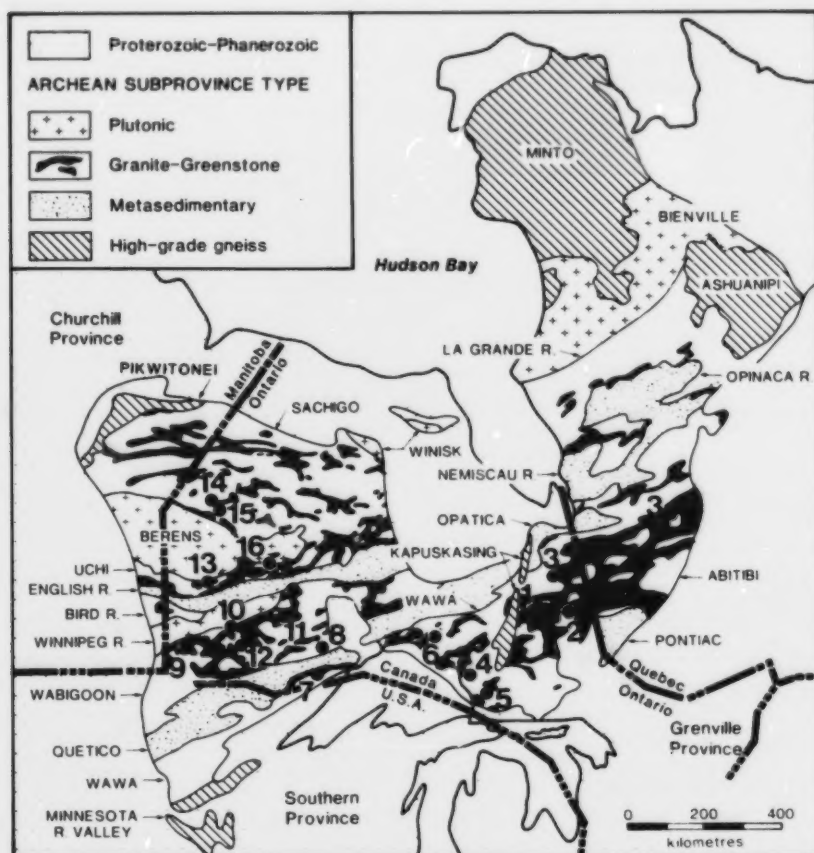


Figure 1.4. Sample locations for U-Pb zircon and titanite geochronology in greenstone terranes of the Superior Province. All determinations with the exception of Mortensen (1987) were carried out at the Jack Satterly Laboratory, Royal Ontario Museum, Toronto.

Figure 1.4 shows the locations of samples dated, and Table 1.2 lists the range of volcanic and plutonic ages obtained from greenstone belts in the Superior Province where U-Pb zircon geochronology has been completed. The data are considered in the following discussion in the context of three major chronographic zones: the northwestern Superior Province, the Wabigoon Subprovince, and the southeastern Superior Province. This discussion summarizes the Archean igneous activity in the Shield, and is extrapolated from all the published U-Pb zircon and titanite ages in Ontario (Figure 1.5).

The greenstone terranes of the northwestern Superior Province contain the oldest volcanic rocks dated in the Province (3013 ± 10 Ma.). Volcanic episodes were separated by long intervals of no magmatic activity. For example, three distinct felsic volcanic episodes occurred in the Red Lake belt: two major episodes, at 3000 to 2900 Ma. and 2760

to 2730 Ma., and a minor episode at 2830 Ma. (Corfu and Andrews 1987).

The greenstone terranes of the southeastern Superior Province contain younger volcanic rocks (2750 to 2700 Ma.) than those of the northwestern terranes, with the exception of one felsic volcanic unit at Hemlo which returned an age of 2772 Ma. (Corfu and Muir 1988). Samples from the southeastern terranes also indicate short time intervals between volcanic episodes; for example, in the Timmins area, three distinct felsic volcanic episodes occurred at 2727, 2714 and 2700 Ma. (Marmont and Corfu 1988).

Volcanism in the Wabigoon Subprovince is generally similar in age to that of the southeastern greenstone terranes (2755 to 2700 Ma., Davis and Edwards 1986), but some greenstones in the Wabigoon (eg. the Lumby Lake area) are as old as 2999 to 2990 Ma. (Davis and Jackson 1985).

PART I. DEPOSITIONAL MODEL

AREA	RANGE OF AGES	
	VOLCANIC	PLUTONIC
SOUTHEASTERN GREENSTONE TERRANES		
1. Timmins	2730 - 2698	2691 - 2673
2. Kirkland Lake	2747 - 2701	
3. Detour Lake - Chibougamau	2802 -	2722 - 2697
4. Gamitagama	2713 -	2668 -
5. Batchawana	2730 - 2701	2678 - 2668
6. Hemlo	2772 - 2695	2690 - 2670
7. Shebandowan	2733 - 2689	2684 -
WABIGOON GREENSTONE TERRANE		
8. Lumby Lake	2999 - 2990	2929 -
9. Kakagi	2731 -	2709 - 2699
10. Savant - Crow Lakes	2755 - 2703	2695 -
11. Sturgeon Lake	2733 - 2717	
12. Manitou Lake	2755 - 2703	2695 -
NORTHWESTERN GREENSTONE TERRANES		
13. Red Lake	2989 - 2744	2719 - 2700
14. Favourable Lake	2918 -	2732 - 2710
15. North Spirit Lake	3023 -	2744 -
16. Uchi - Confederation Lakes	2959 - 2738	

Table 1.2. Summary of U-Pb zircon and titanite geochronological determinations for volcanic and plutonic rocks from greenstone terranes of the Superior Province. All determinations, with the exception of Mortensen (1987), were carried out in the Jack Satterly Laboratory, Royal Ontario Museum, Toronto. Sources of data include: Timmins - Marmont and Corfu (1988), Frarey and Krogh (1986); Kirkland Lake - Marmont and Corfu (1988); Detour Lake - Chibougamau - Marmont and Corfu (1988), Mortensen (1987); Gamitagama - Krogh and Turek (1981); Batchawana - Corfu and Grunsky (1987); Hemlo - Corfu and Muir (1988); Shebandowan - Corfu and Stott (1986); Lumby Lake - Davis and Jackson (1985); Kakagi - Davis and Edwards (1986); Savant - Crow Lakes - Davis et al. (1981); Sturgeon Lake - Davis and Trowell (1982); Manitou Lake - Davis et al. (1981); Red Lake - Corfu and Andrews (1987); Favourable Lake - Corfu et al. (1981); North Spirit Lake - Corfu and Wood (1986); Uchi - Confederation Lakes - Nunes and Thurston (1980).

Across the Superior Province, many intrusions have yielded ages similar to their volcanic counterparts. In addition, a number of younger intrusions which cut the supracrustal package have been dated; in most cases, field relationships indicate that the supracrustal rocks have been folded and/or tilted prior to the emplacement of these intrusions. These intrusions are the youngest host lithology to gold mineralization (see Chapters 4 and 5). In the northwestern greenstone terranes, they

have ages between 2720 and 2700 Ma. (Corfu and Andrews 1987), and in the southeastern terranes, they range in age between 2690 and 2665 Ma. (Corfu and Stott 1986, Corfu and Grunsky 1987, Marmont and Corfu 1988).

In the Timmins area, these intrusions comprise two suites (Marmont and Corfu 1988): a quartz-bearing suite (usually quartz porphyries) which dates between 2691 and 2688 Ma., and a sil-

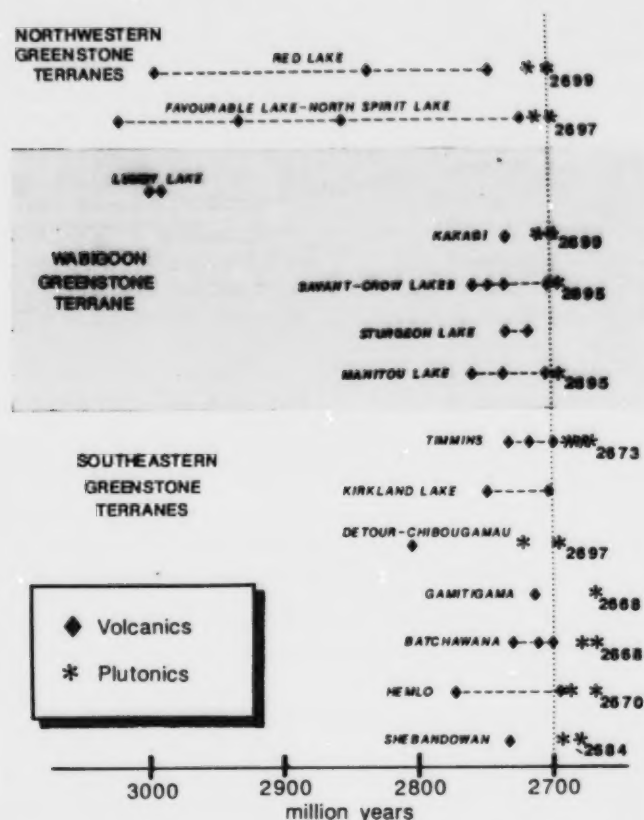


Figure 1.5. Summary of geochronological data for Archean volcanism and plutonism in the Superior Province. See Table 1.2 for additional information.

ica-undersaturated suite (monzonite, syenite, lamprophyre?) which dates between 2678 and 2673 Ma. The significance of these intrusions to gold mineralization is discussed in detail in Chapter 4, Part I of this volume.

In summary, the geochronological data indicate that the latest volcanic activity occurred at different times in the greenstone belts across the Superior Province (Figure 1.5). This final volcanic activity is older in the northwestern terranes (2740 to 2720 Ma.) than in the southeastern terranes (approximately 2700 Ma., ignoring the one anomalously old sample from the Hemlo area). The secular variation reflected in the age of the late Archean plutons also implies that cratonization of the Superior Province occurred earlier in the northwestern terranes than the southeastern terranes.

At present, U-Pb geochronology of the high grade metamorphic terranes is restricted to but a few parts of the Superior Province (Percival and Krogh 1983, Krogh *et al.* 1988, Corfu 1987). Corfu

(1987) demonstrated that titanites (used to obtain metamorphic and metasomatic ages) from the volcanic units dated by zircons (used to obtain formational ages) yield identical or slightly younger ages than are obtained from the zircons. This suggests that the greenstone terranes had a rapid cooling history and were not disturbed after cratonization. Corfu (1987) also showed that both zircons and titanites from granulite terranes in the Superior Province yield multiple ages, which suggests a protracted thermal and magmatic history. This aspect of late Archean tectonism is fully discussed in Part II.

1.5. LITHOLOGICAL ASSOCIATIONS OF GOLD DEPOSITS

All Archean rock types within the greenstone belts of the Superior Province are known to host gold mineralization. The metamorphic grade of these rocks ranges from sub-greenschist to middle amphibolite facies. As well, individual gold deposits

PART I. DEPOSITIONAL MODEL

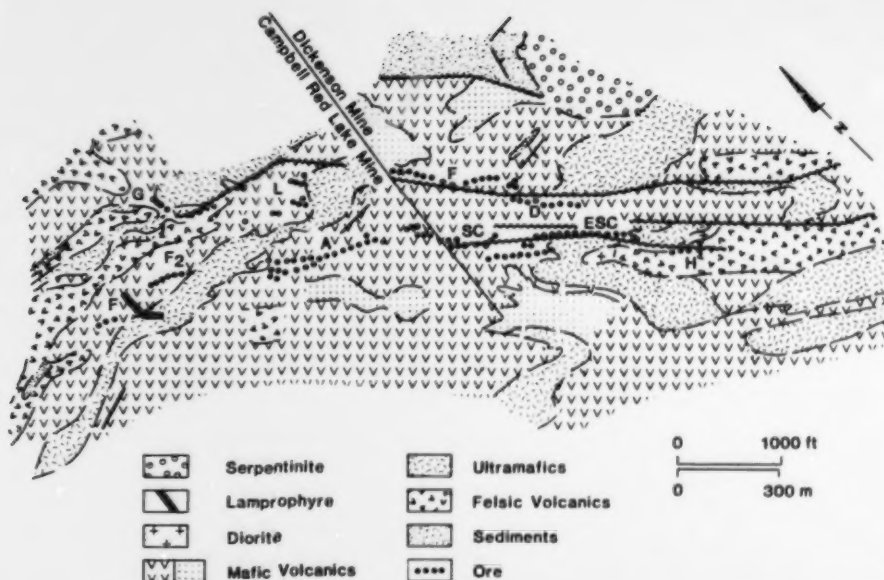


Figure 1.6. Generalized geology and ore zones of the 15th level at the Campbell Red Lake and Dickenson (A.W. White) Mines, Red Lake. Diagram after Lavigne and Crocket (1983).

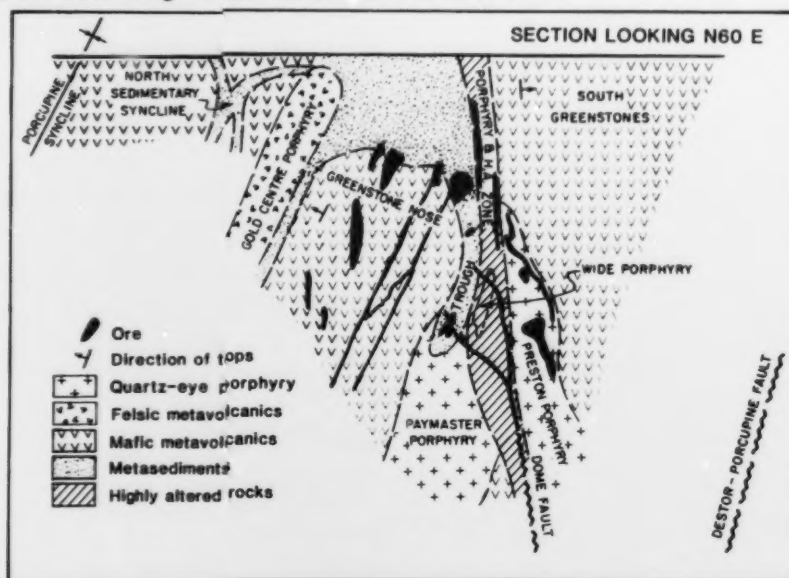


Figure 1.7. Generalized geological cross-section of the Dome Mine, Timmins. Diagram modified after Girdwood et al. (1983).

can be contained in a number of lithologies. The Campbell Red Lake and Dickenson (A.W. White) Mines in the Red Lake camp (Figure 1.6), and the Dome Mine in the Timmins camp (Figure 1.7) are good examples of this heterogeneity of host lithologies in a single deposit. In these mines, host rocks include ultramafic, mafic and felsic metavolcanics, clastic metasediments, and mafic to felsic intrusions.

Previous models for the generation of lode gold deposits have suggested that particular lithologies in greenstone belts are preferred hosts to mineralization. One test of this hypothesis is a comparison of the frequency of occurrence of gold mineralization in a lithology, at a gold camp or deposit scale, with the frequency of occurrence of that lithology in a greenstone belt. The results of this

comparison for the Abitibi greenstone belt, done by Colvine *et al.* (1984), are shown on Figure 1.8. The data permit several observations to be made about the occurrence of gold mineralization, and also clarify some often quoted concepts about gold deposits:

- (1) Mafic and ultramafic volcanic and intrusive rocks have been suggested as preferred host rocks to gold. While these lithologies are widespread within greenstone belts, their frequency of occurrence as hosts to mineralization is less than their belt-wide occurrence.
- (2) Clastic metasediments host mineralization in a frequency approximately proportional to their belt-wide occurrence.
- (3) Felsic metavolcanic rocks have a somewhat higher incidence in the mineralized areas than within greenstone belts as a whole. This may be an artifact of lithological identification: in some deposits, felsic intrusive units have been identified as felsic

metavolcanic rocks, due to destruction of intrusive contacts by deformation. (eg. the Pearl Lake Porphyry in the McIntyre Mine, Davis and Luhta 1978). In other areas, highly silicified and/or feldspathized mafic metavolcanic rocks have been misidentified as felsic metavolcanic rocks (eg. the Ross Mine, Troop 1986).

- (4) Chemical metasediments such as banded iron formations are only a minor proportion of greenstones. The observation in many greenstone terranes that these lithologies, when present, constitute a favourable chemical and structural host to mineralization is supported by the data on Figure 1.8. However, they do not represent a major ore host. Hodgson (1983a) demonstrated that, in fact, banded iron formations are more frequently associated with economically poorer deposits.

- (5) The post-volcanic felsic plutons within the Abitibi belt are minor in volume in comparison with syn-volcanic felsic intru-

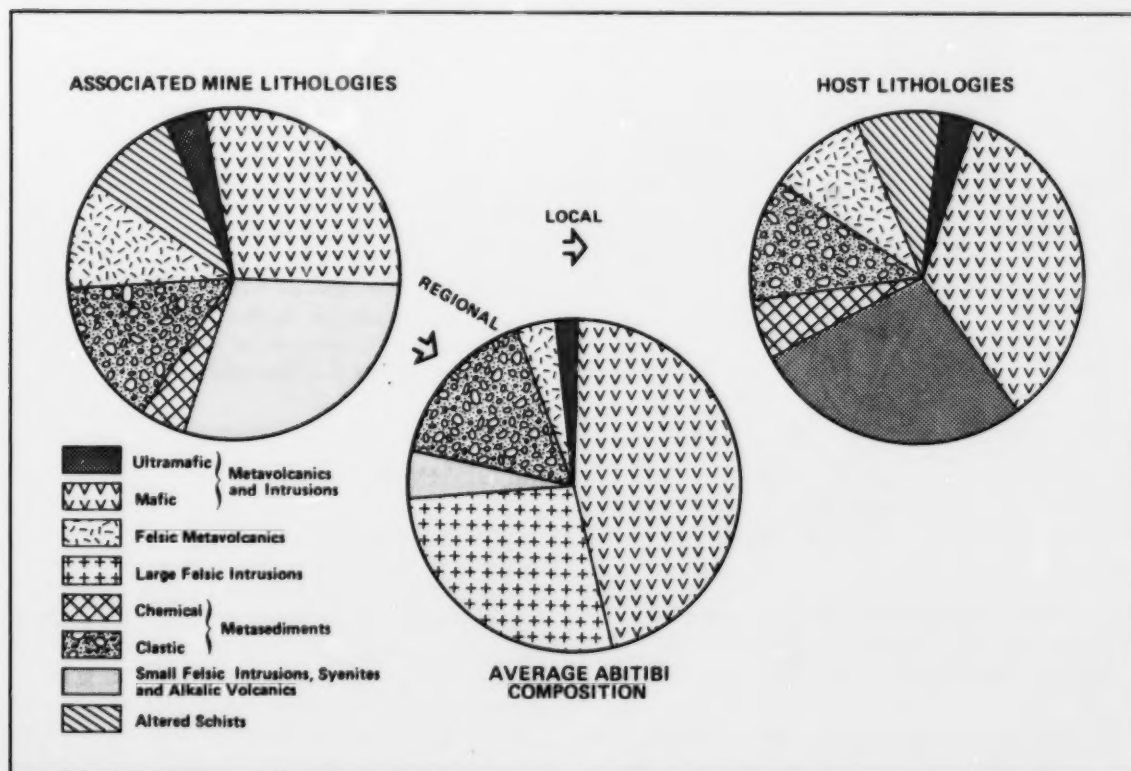


Figure 1.8. Relative abundances of greenstone lithologies present: (1) in the Abitibi greenstone belt, (2) associated with gold mineralization at a property scale, and (3) hosting gold mineralization at an orebody scale. Diagram after Colvine *et al.* (1984).

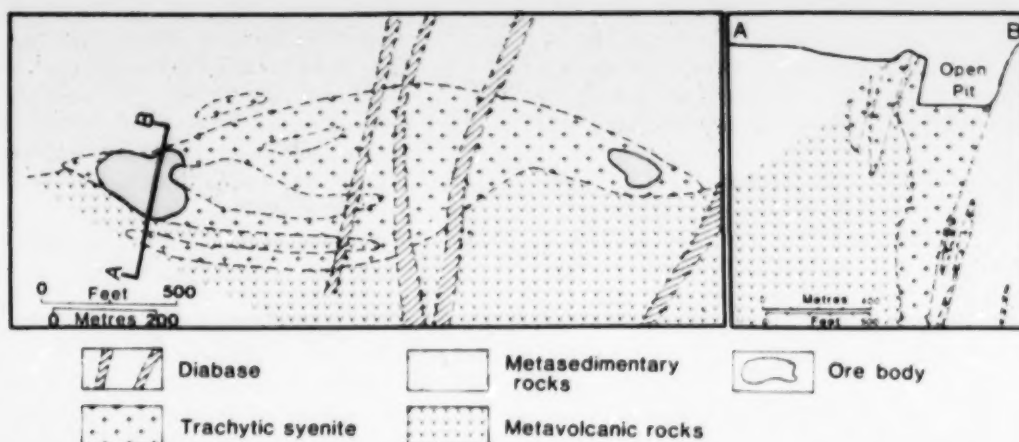


Figure 1.9. Generalized geology of the Young Davidson Mine, Matachewan. Diagram after Sinclair (1982).

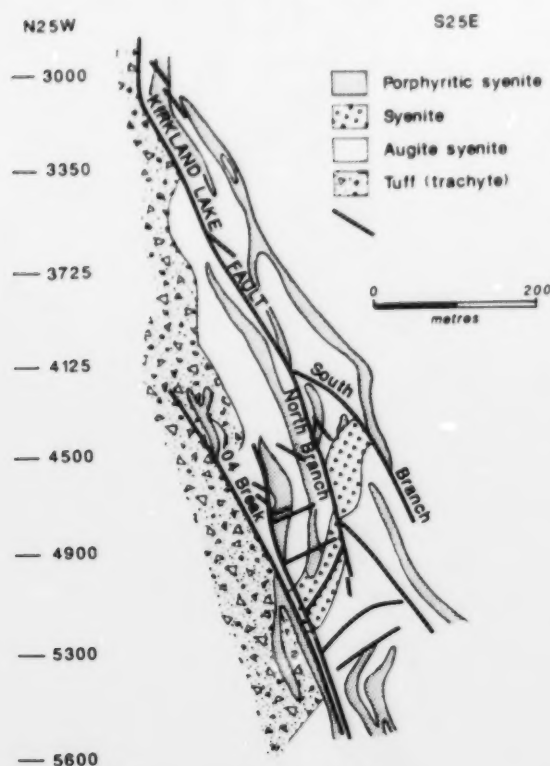


Figure 1.10. Generalized geological cross-section through the No. 1 Shaft of the Macassa Mine, Kirkland Lake. Diagram after Charlewood (1964).

sions, constituting less than 4% of the belt. However, these post-volcanic intrusions are very common in mineralized areas, constituting more than 25% of host rocks to mineralization. In several deposits, gold miner-

alization is either completely (eg. the Young-Davidson Mine at Matachewan) or predominantly (eg. the Macassa Mine at Kirkland Lake) hosted by felsic to intermediate, silica-saturated to undersaturated intrusions which cut the folded and tilted supracrustal package. In the Young-Davidson Mine (Figure 1.9), trachytic syenite intruded at the contact between mafic metavolcanic rocks and clastic metasediments is the only host to mineralization (Sinclair 1982). In the Macassa Mine (Figure 1.10), a suite of mafic to felsic syenites intrudes the supracrustal rocks, which include Timiskaming-type sediments and volcanics, and is host to most of the gold mineralization. The role of felsic magmas in the development of Archean gold deposits is discussed in more detail in Chapter 4, Part I and in Part II of this volume.

1.6. SUMMARY

For the most part, this chapter has presented field observations about Archean lode gold deposits. These field data were supplemented by precise geochronology, which establishes an absolute time frame in which to discuss both the evolution of the greenstone terranes which host the gold deposits, and the probable time of gold mineralization. Collectively, the field and geochronological data are consistent with the formation of the gold deposits after the intrusion of late tectonic plutons into the folded and tilted supracrustal sequences of the greenstone belts.

In most geological and mineral deposits studies, cross-cutting relationships similar to the ones documented in the preceding sections would be accepted as conclusive evidence of a sequence of events. A simplistic analogy would be the interpretation of the age of a dike swarm cutting across folded rocks. However, gold deposition may not have been so simple a process, and a more thorough examination of the Archean gold deposits may be necessary to develop a comprehensive understanding of how they formed.

In the remaining chapters of Part I of this volume, the hypothesis that gold was the product of a major mineralizing event in the Archean is examined in the context of its structural setting, associated wall rock alteration and relationship to metamorphism, and association with felsic magmatism. These data, complemented with geochronology and studies to characterize gold-bearing fluids, are used to construct a depositional model for Archean lode gold deposits, correlated in time across large areas of the Superior Province.

PART I. DEPOSITIONAL MODEL

2. STRUCTURAL SETTING

2.1. INTRODUCTION

It has long been known that Archean lode gold deposits occur in deformed rocks (cf. Hurst 1935, Gunning and Ambrose 1937, Newhouse 1940). At the largest scale, gold deposits are clearly associated with shear systems, herein called deformation zones. These deformation zones are mappable features on a regional scale. As shown on Figure 1.2, all of the major gold camps in the Superior Province of Canada, including Rice Lake, Red Lake, Beardmore - Geraldton, Hemlo, Wawa, Timmins, Kirkland Lake, Val d'Or - Malartic, Casa Berardi and Duparquet, are associated with deformation zones which are of regional extent. It seems probable, given this strong correlation, that the deformation zones played an important, if not fundamental, role in the formation of the gold deposits.

Within certain portions of the regional deformation zones, smaller scale structures can be shown to have played a direct role in localizing gold deposits. Deposits as a whole, and individual orebodies and ore shoots at a smaller scale, commonly fill specific structural features, such as fractures, shear surfaces, extension cracks, fold hinges and intersections of planar structures. These depositional sites within deformation zones typify sites that were dilatant and permeable during deformation. Their occurrence and orientation, which often seem random upon cursory inspection, can generally be explained by strain analysis at different scales in the deformation zones.

In addition, there is evidence, from a composite analysis of the different gold deposits, that these structural sites of gold deposition change in form with depth. Therefore, the geometry of gold deposits at all scales changes in style with depth of formation. This implies a vertical zonation to the geometry of gold deposits that provides an element of predictability, fundamental to an applied model for Archean gold deposits.

This chapter addresses the structural controls of gold mineralization, by relating the mineralization to successively smaller structures. The regional deformation zones are first described, and their development is discussed. The structural controls of both gold deposits and ore veins and shoots are then described in light of field observations in gold deposits across Ontario, and a model proposed.

Finally, from all of this evidence, the geometry of gold deposits in these deformation zones is

shown to be predictable if a model of vertical zonation which represents depth of burial beneath the paleosurface during mineralization is applied

2.2 DEFORMATION ZONES

The deformation zones or shear zones associated with Archean lode gold deposits are zones of anomalously high strain. Not only can these zones be recognized at the Shield scale (Figure 1.2), but they can be clearly delineated within and adjacent to individual greenstone belts (Figures 2.1, 2.2, 2.3). They transect and deform all Archean lithologies and hence are not restricted to certain units. Nevertheless, these zones form discrete, linear, mappable units of deformed rock up to several kilometres in width, and up to hundreds of kilometres in length.

These linear units of deformed rock occur in generally predictable orientations in the Canadian Shield (Figure 1.2) and are located in certain preferred settings, as noted in Section 2.2.1 below. The deformation zones record a history dominated by progressive simple shear, as reflected by the pronounced asymmetry in the local structures. These enclosed structures and the boundary relationships of deformation zones provide some basic field characteristics, which are summarized in Section 2.2.2.

2.2.1. REGIONAL SETTING OF DEFORMATION ZONES

The regional - scale deformation zones transect the boundaries of individual greenstone belts or form the boundary between granite - greenstone and metasediment - dominated domains (Figure 1.2). In many cases, it is impossible to correlate lithologies and structures across such zones, and movement across them is uncertain. For example, Bau (1979) suggested a right lateral offset of over 120 km along the Quetico Fault, which forms the boundary between the Quetico and Wabigoon Subprovinces.

A compilation of the regional - scale deformation zones in the Superior Province (Figure 1.2) indicates consistencies in both orientation and displacement sense: northeasterly - trending zones show left - handed and commonly reverse sense of displacement; southeasterly - to easterly - trending zones are right - handed. Recent work suggests that some of the easterly - trending zones also record vertical displacement (Smith and Thomas 1986, Sanborn 1986, Williams 1986, 1987). Overall, the orientation and kinematics of these deformation

PART I. DEPOSITIONAL MODEL

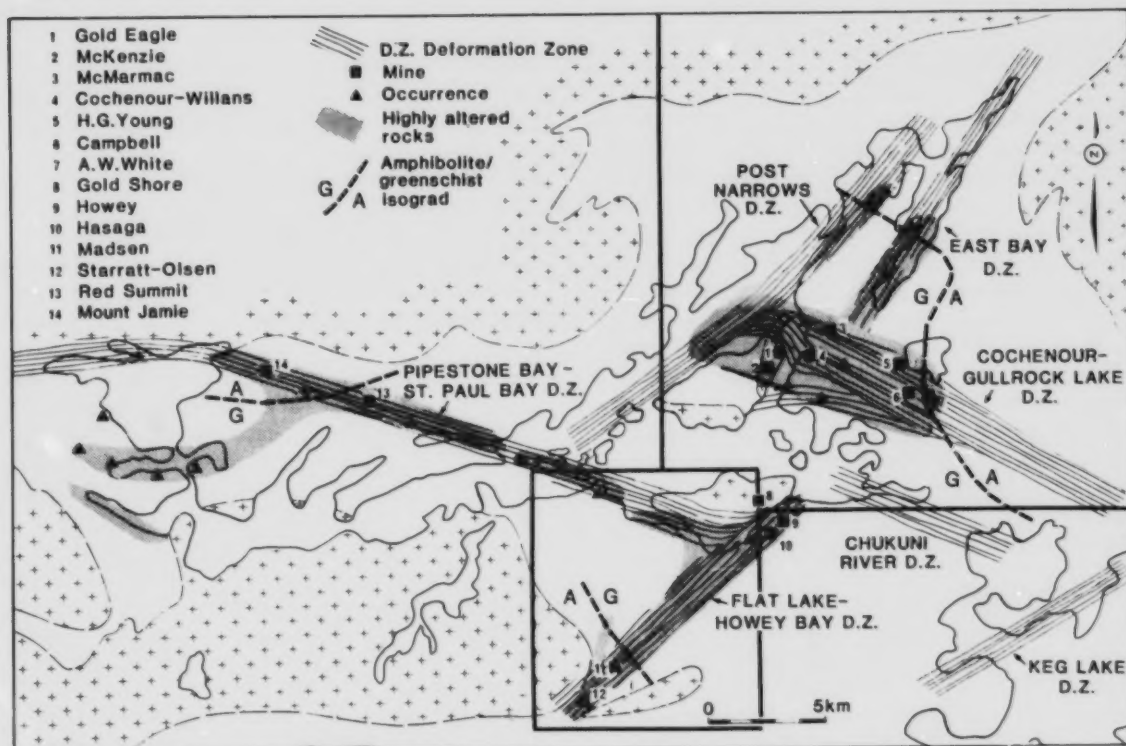


Figure 2.1. Distribution of gold deposits within highly altered deformation zones in the Red Lake area (after Andrews 1986).

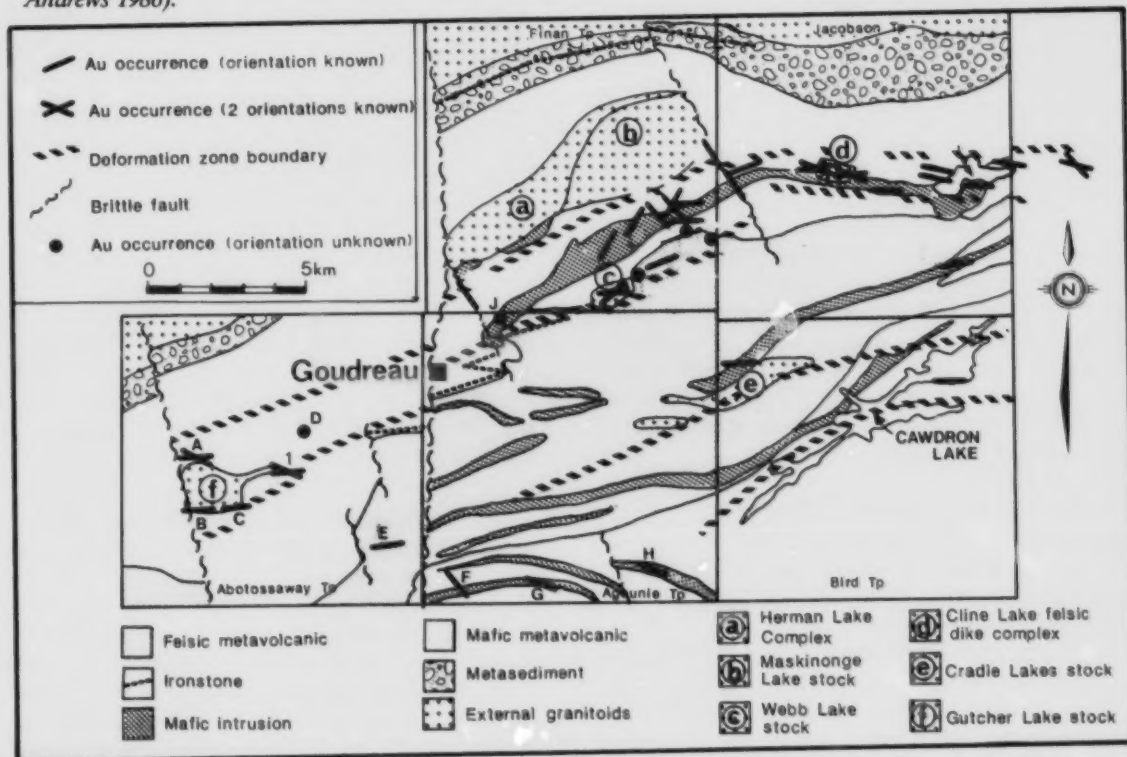


Figure 2.2. Distribution of gold mineralization within the Goudreau Lake Deformation Zone in the Wawa area (after Heather and Arias 1987).

zones reflect a regional north - northwesterly - directed compression.

These consistent orientations provide an element of predictability. In addition, the faults and shear zones preferentially, and thus predictably, form along east - west-trending subprovince boundaries, northwest-trending greenstone - granite contacts (eg. in Uchi Subprovince - Stott and Wallace 1984) and within linear greenstone belts that trend both north-east and northwest (eg. Pipestone - Cameron Lake Fault - Wabigoon Subprovince and the Casa Berardi zone in the northern Abitibi belt of Quebec).

Deformation zones also form in response to the ascent and emplacement of both major granitoid batholiths that are external to greenstone belts, and smaller plutons that occur within greenstone belts. These zones are discussed briefly by Stott and Smith (1988), who recognized that gold - bearing structures in the Lake of the Woods and Pickle Lake areas were developed in deformation zones that had been created or reactivated by the emplacement of plutons. Recent work in the Uchi and Wabigoon Subprovinces (Andrews *et al.* 1986, Smith 1987) indicates that significant deposits, such as those in Red Lake, are hosted in regional deformation zones, but

that ore zones are controlled by structures indicative of vertical movement related to the emplacement of adjacent batholiths, rather than by structures resulting from transcurrent displacement.

Thus, two regional settings are evident for deformation zones that are associated with gold deposits. One results from large scale crustal movement or shortening; the other results from emplacement of plutons and batholiths. In general, zones which resulted from the first process have greater strike length and host most of the significant gold deposits.

2.2.2 FIELD CHARACTERISTICS OF DEFORMATION ZONES

Deformation zones constitute mappable units on the basis of the criteria summarized below, which are diagrammatically shown on Figure 2.4.

1) Lithostratigraphic units are characteristically transposed into parallelism with deformation zones. Lithologies may be rotated (A), folded (B), dislocated (C), truncated (D), thinned (E), thickened (F), repeated (G), or transposed (H) (see Figure 2.4) within or along the margins of deforma-

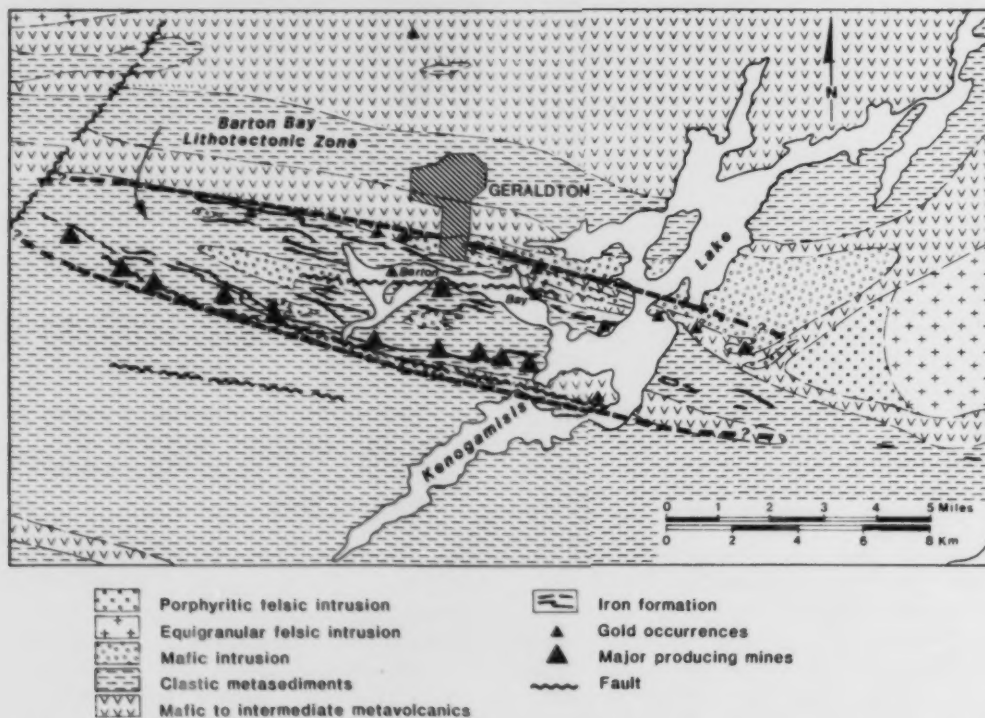


Figure 2.3. Distribution of gold mineralization within the Barton Bay Lithotectonic Zone in the Geraldton area (after Macdonald 1984).

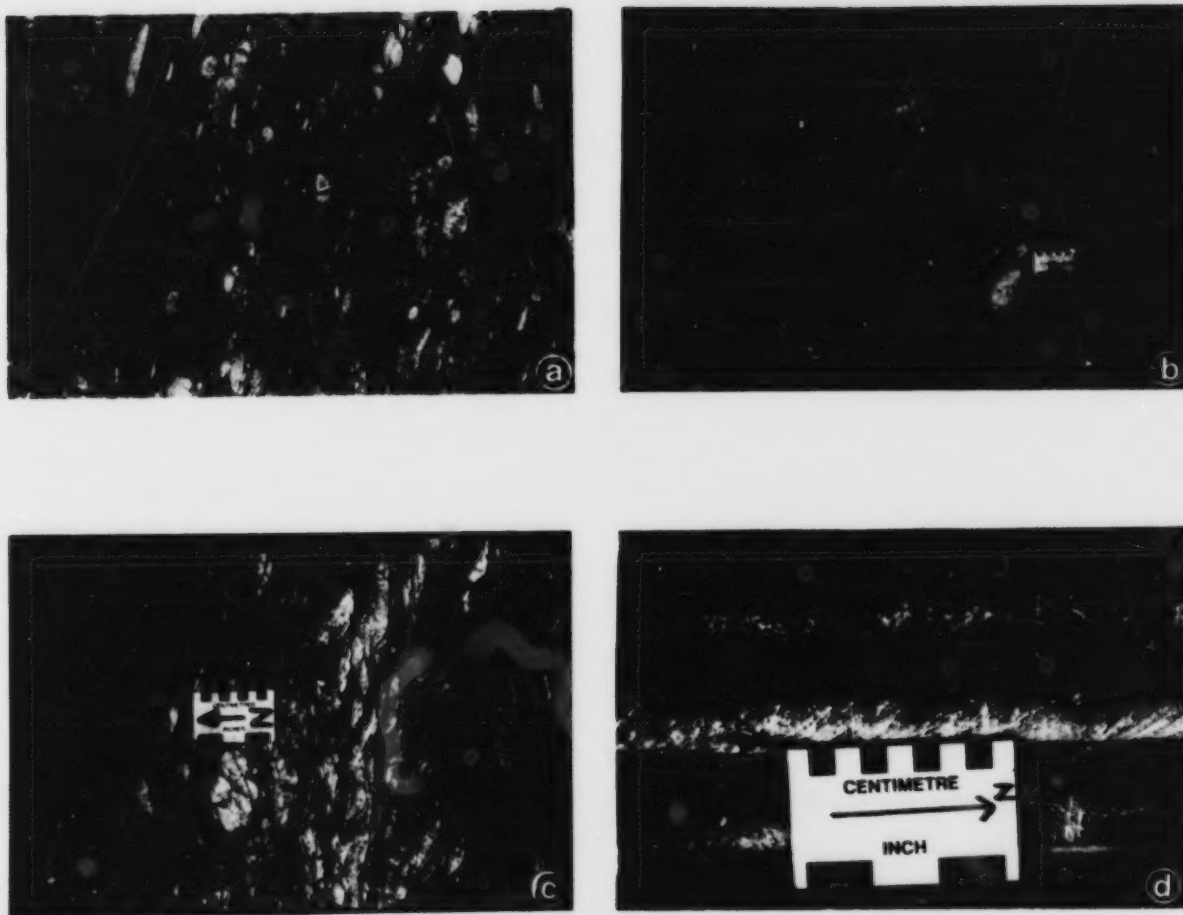


Plate 2.1. (a) Slightly deformed conglomerate; clasts of volcanic and sedimentary material are elongate, while granitic clasts are not. From Wawa, Ontario.
 (b) Highly deformed conglomerate; clasts of volcanic material are highly elongate, quartz veins and slightly elongate granitic clasts are rotated and transposed. From Wawa, Ontario.
 (c) Intensely deformed conglomerate; mylonitic layering is cut by a secondary cleavage and deformed by kink bands. Granitic clasts are broken and offset. From Wawa, Ontario.
 (d) Progressive deformation of coarse grained basalt from weakly foliated material at top, to talc-chlorite schist at base. From Shoal Lake Deformation Zone, Ontario.

tion zones. On an outcrop scale, and in some instances on a mine scale, deformation zones are parallel to bedding and are only evident as anomalous reversals in facing direction (I; Figure 2.4).

2) Deformation zones are commonly discordant to lithological boundaries on a regional scale and therefore contain deformed equivalents of various rock types that typify greenstone - granite terranes. Within deformation zones, strain intensity is highly variable; lozenge - shaped lithons and enclaves of less deformed rock are common (J; Figure 2.4), and a particular protolith can often be traced into its mylonitized equivalent (Plate 2.1).

3) Deformation zones contain a variety of fault rocks, such as gouge, mylonites and phyllonites (Figure 2.5), and foliated rocks, including chlorite and sericite schists.

4) Deformation zones frequently consist of many individual sheared zones that are rarely linear and continuous, but which anastomose, bifurcate, or are discontinuous in both vertical and horizontal dimensions.

5) Schistosity is subparallel to the trend of a deformation zone where strain is most intense and may be oblique to it along its margins, depending

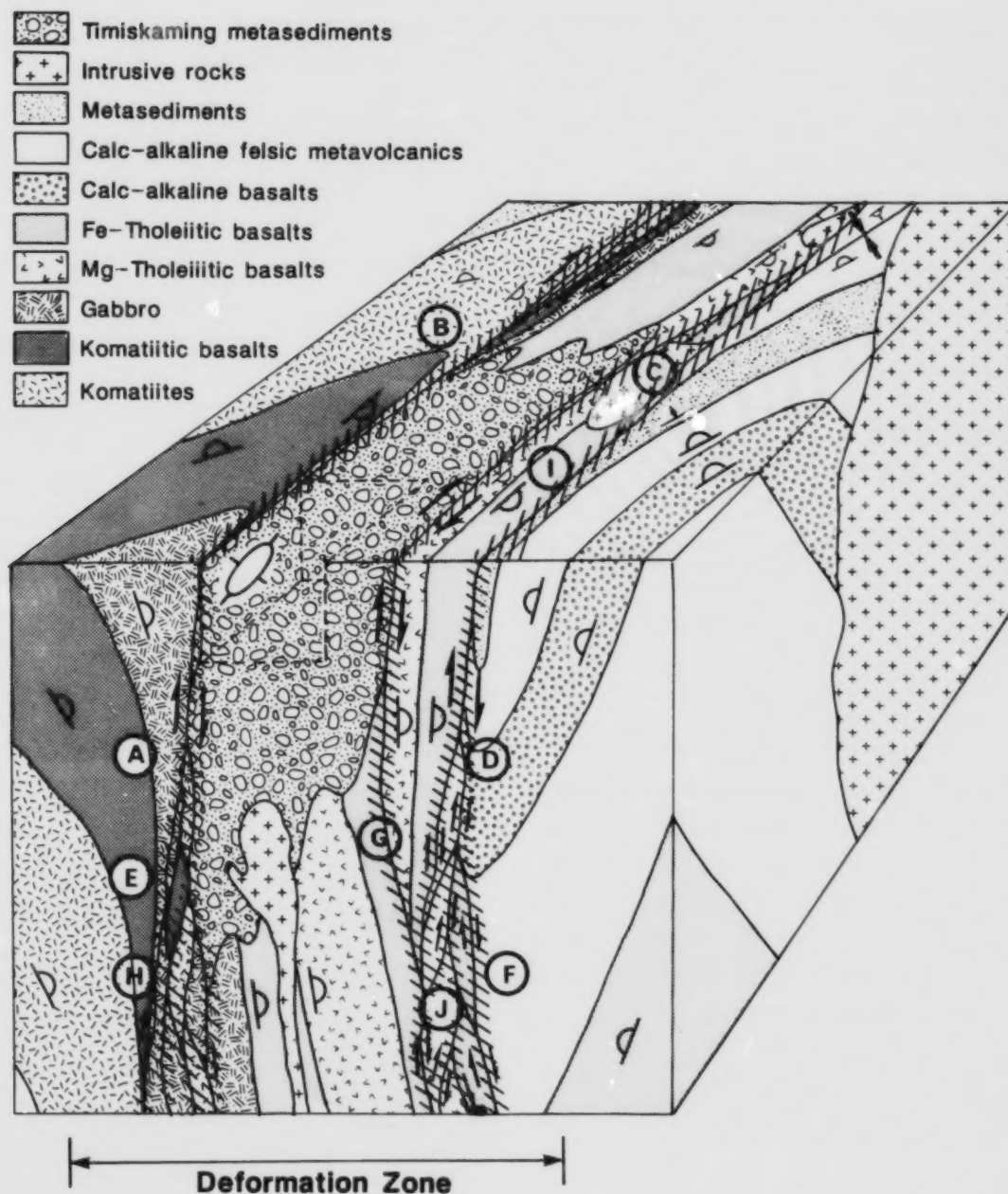


Figure 2.4. A schematic block diagram indicating large-scale features of deformation zones. Letters refer to structural features detailed in Section 2.2.2.

on the orientation of the slip direction. Planar fabrics, which may be superimposed on one another, may represent temporally discrete or contemporaneous deformation events (conjugate foliations, kink bands, C - S structures and shear bands (Plate 2.2). Conjugate cleavages and S fabrics are most common in the least strained, while C and C' fabrics (Berthé

et al., 1979) are common in the most intensely strained rocks. Stretching and mineral lineations grew along, or rotated to, a direction subparallel to the slip direction. Intersection lineations, such as between C and C' fabrics, tended to form approximately orthogonal to the slip direction.

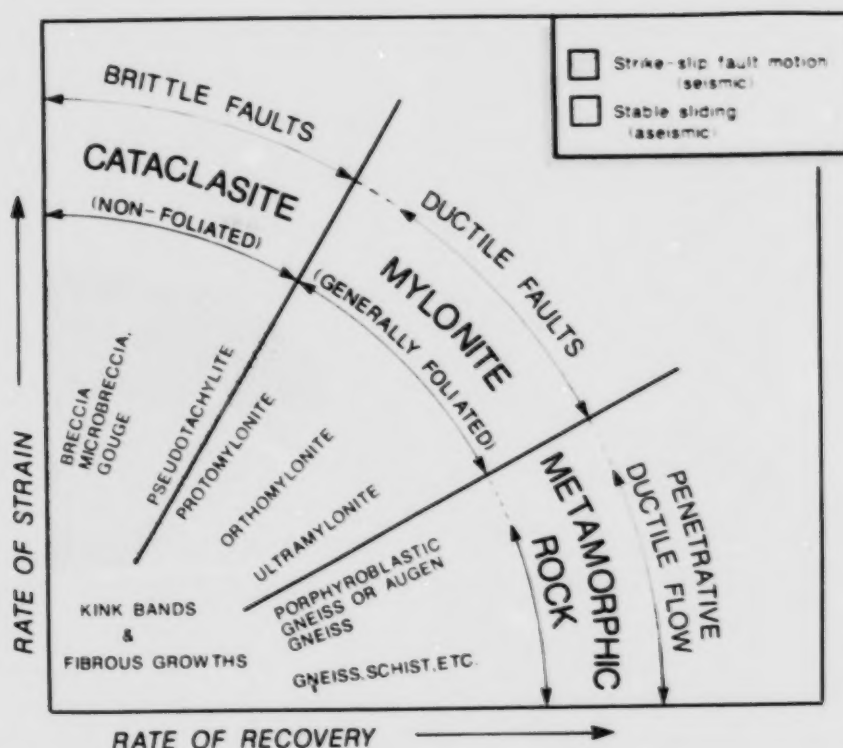


Figure 2.5. Rock types found within deformation zones, indicating their relationships to qualitative strain and recovery rates.

6) Progressively deforming zones may contain schistosomes which are themselves folded (Plate 2.2f). The axial orientations of these folds are variable, but with increasing strain they tend towards parallelism with the slip direction. In some cases, sheath folds are formed (Cobbold and Quinquis 1980, Hudleston 1986).

7) Many deformation zones are characterized by structures indicative of both ductile and brittle deformation, as documented for the Cochenour - Gullrock Lake Deformation Zone at Red Lake (Andrews *et al.* 1986) and for the Goudreau Lake Deformation Zone in the Wawa greenstone belt (Arias and Heather 1987). Brittle and ductile deformation may be coeval, or may have occurred at different times and thereby superimposed one on the other (cf. Ramsay 1980).

8) Deformation zones generally contain highly altered rocks. Carbonate and hydrous minerals, as well as quartz and quartz - carbonate veins, are indicative of relatively low P-T conditions. Aluminosilicate and garnet alteration assemblages are present at higher P - T conditions (see Chapter

3 for a detailed discussion of alteration associated with gold deposits). Vein systems in deformation zones are usually complex, or crack - seal type (Ramsay and Huber 1983).

9) Alluvial - fluvial sequences of immature sedimentary and intercalated volcanic rocks, often referred to as "Timiskaming - type", are commonly associated with regional deformation zones (e.g. Corfu and Stott 1986), unconformably overlying older terranes.

10) Syn-to late tectonic, intermediate to felsic, silica-saturated and -undersaturated dikes and stocks have commonly intruded the deformation zones. Lamprophyre dike swarms are also common (Wyman and Kerrich 1986, 1987); these rocks are weakly to strongly affected by shear deformation, but typically behave more competently than their host rocks.

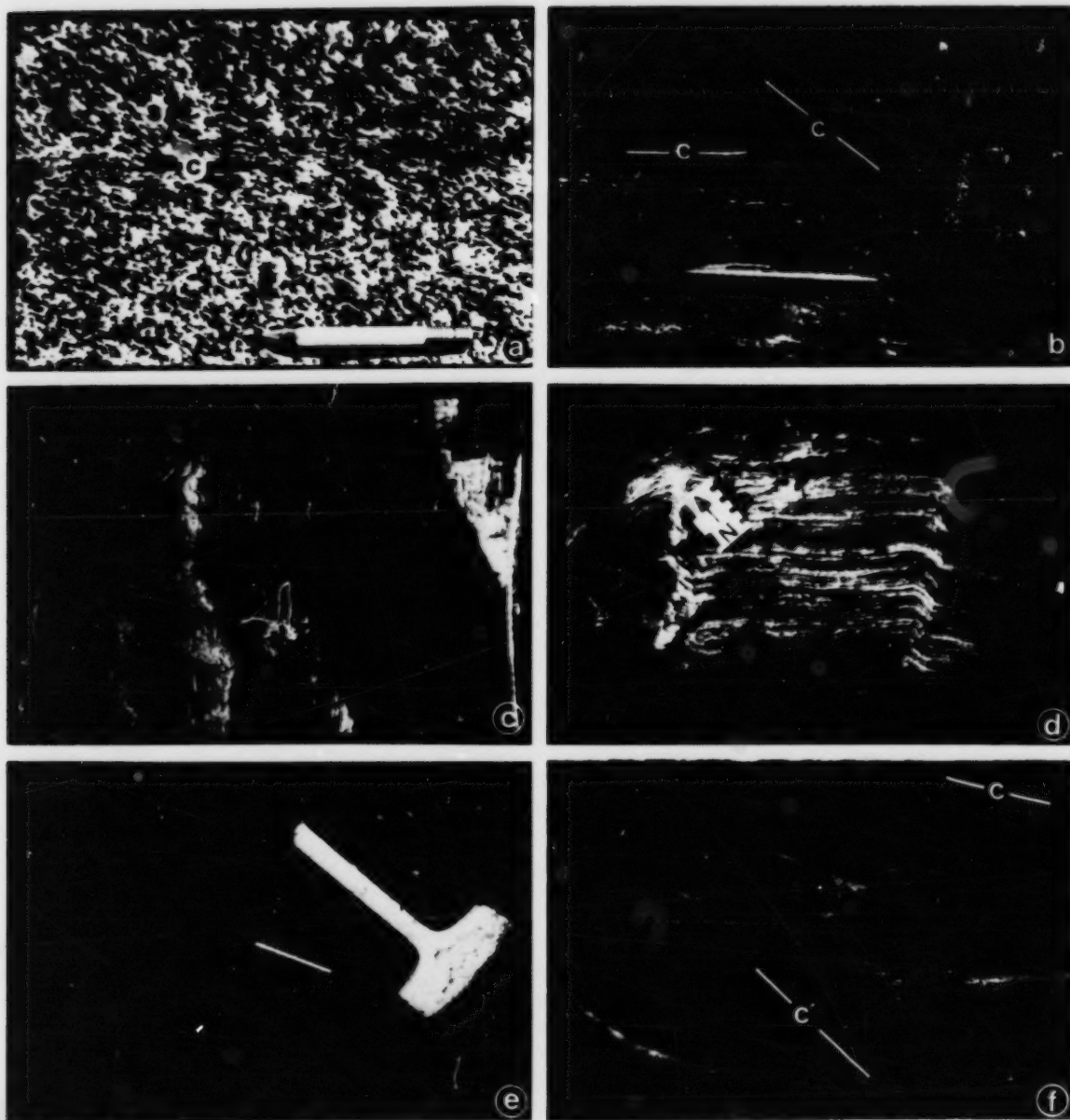


Plate 2.2. (a) Shear zone in gabbro; rotation of (S) fabric towards (C) in centre of zone. From Big Duck Lake, Ontario.
 (b) Well developed crenulation fabric (C') discordant to mylonitic (C) fabric. From Barton Bay Lithotectonic Zone, Ontario.
 (c) Kink bands in ultramylonite from Crowduck Lake-Witch Bay Shear Zone, Ontario.
 (d) Kink bands in ultramylonite at Wawa, Ontario.
 (e) Conjugate foliation (shear bands) in Crowduck Lake-Witch Bay Shear Zone, Ontario.
 (f) Intrafolial fold developed in mylonitized gabbro; C' fabric cuts the fold.

PART I. DEPOSITIONAL MODEL

2.2.3. TIMING AND DEPTH VARIATIONS OF DEFORMATION ZONES

2.2.3.1. Timing

Previous structural studies and modelling of Archean greenstone belts have emphasized the role of gravitational subsidence in the tectonic evolution of Archean crust (eg. West 1980, Dixon and Summers 1983). However, workers in the Superior Province recognize that the Archean deformation history is characterized by a period of vertical motion tectonics and a period of horizontal motion tectonics (Schwerdtner *et al.* 1979). Recently, Stott (1986) and Sanborn (1986) have demonstrated that the stage of horizontal motion tectonics affected wide segments of Archean crust. The question recently being pursued by these workers is whether the horizontal compression that is so widely recorded is a singular event or a series of progressive events.

Recent U-Pb zircon geochronological work at the Royal Ontario Museum demonstrates that the north-northwesterly directed compression is a sequence of separate late stage shortening events that are progressively younger in age towards the southern part of the Superior Province (Stott *et al.* 1987, Corfu and Stott 1986, Davis *et al.* 1986, Corfu 1988). The regional scale fault zones and shear zones are culminating stages in this southward

progressive history of crustal shortening. Accordingly, it is concluded that the gold deposits hosted within these late stage deformation zones are an integral part of a culminating stage in Archean crustal evolution.

2.2.3.2. Depth Variation

Studies of deformation mechanisms and the physical properties of rocks and minerals suggest that deformation zones should change their character with depth in the crust. Field study of deformation zones whose present surface exposure allows recognition of an original depth profile have been discussed by Ramsay (1980), Sibson (1977, 1983), and Xu *et al.* (1987), among others. The anticipated changes are illustrated schematically by Figure 2.6.

At shallow depths (upper crust, sub- to lower greenschist facies), deformation zones are narrow, discrete faults that commonly contain lozenges of undeformed rock (lithons) separated by zones of gouge and breccia. At intermediate depths (middle and upper crustal levels, greenschist to lower amphibolite facies), shearing becomes increasingly ductile. As a result, the deformation zone anastomoses and widens, and brittle structures become less common. At great depth, occurrence of lithons diminishes, deformation zones become continuous, and brittle deformation occurs only under conditions of rapid changes in pore fluid pressure or strain

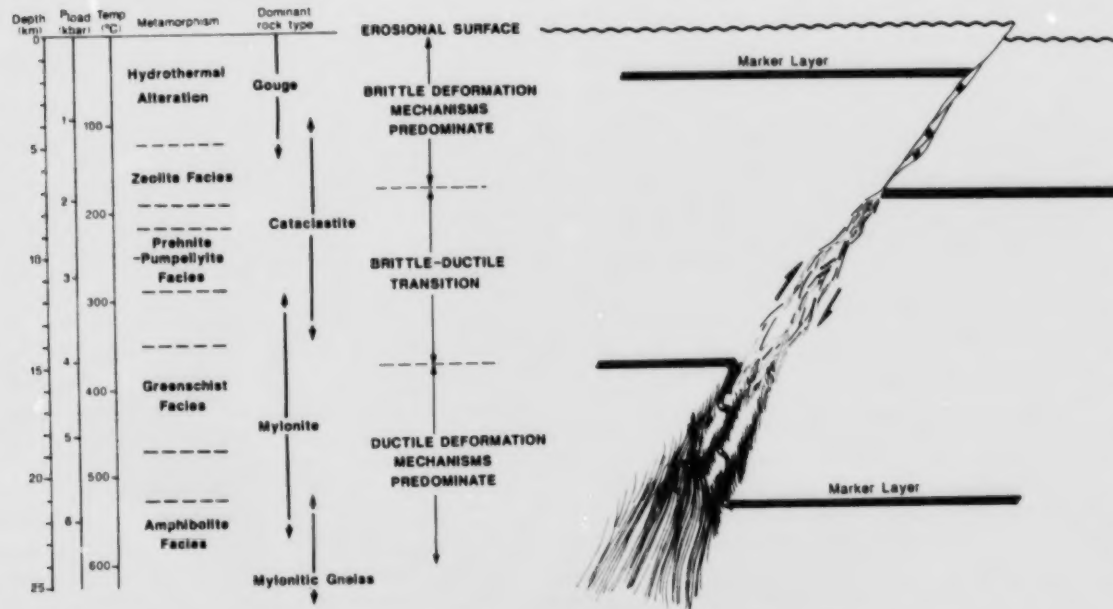


Figure 2.6. Deformation regimes as a function of depth (modified from Simpson 1986, Sibson 1977, 1983).

rate. At these great depths, deformation zones may become shallow - dipping, taking advantage of horizontal anisotropies (Coward 1984).

The presence of a pore fluid in a deforming shear zone will have a considerable effect on the deformation style, and will modify the relative fields illustrated on Figure 2.6.

Brittle deformation may be induced or promoted by high fluid pressures (Price 1966), or by the interaction of fluid with particular lithologies. Introduction of CO_2 , with or without H_2O , affects mineral stabilities and, hence, rock ductility: field evidence indicates that dolomite - rich and/or silicified rocks are relatively resistant to shear, and behave in a brittle manner in comparison to unaltered rocks.

Paradoxically, ductile deformation can also be promoted by high fluid pressure, which lowers the temperature at which grain boundary diffusion occurs, induces pressure solution and hydrolytic weakening, aids in strain - induced recrystallization and neomineralization, and promotes retrogression and alteration.

Since deformation zones are exposed at different erosional levels throughout the Superior Province, variation in the character of shear zones with depth can be related to corresponding observations of systematic variation of small scale structures, which are discussed in the next section.

2.3. CONTROL AND DISTRIBUTION OF GOLD MINERALIZATION

Gold mineralization is hosted in small scale structures within the larger deformation zones. These structures are highly permeable zones, preferred sites for mineralization. As described in the preceding section, a change from brittle to ductile deformation occurs with increasing depth in a deformation zone, as ambient temperature and pressure increase: this change in deformation style is reflected in the form of gold - bearing structures. Evidence presented elsewhere in this volume (see Chapters 1 and 3, Part I) also shows that the gold is associated with large volumes of fluid that have altered the host rocks: this fluid, through its influence on the ductile - brittle behaviour of the rocks, will also affect the form and distribution of gold - bearing structures.

This section examines the distribution and

styles of gold mineralization from camp to vein scale, and links the observed variation to structural controls.

2.3.1. DISTRIBUTION OF GOLD CAMPS

At the largest scale, areas of enhanced permeability along deformation zones may contain a number of significant deposits which together form a gold camp (eg. Timmins, Kirkland Lake, Red Lake).

Colvine *et al.* (1984) suggested that some gold camps are located in portions of deformation zones which have undergone increased extension during wrench faulting. Pull - apart basins developed during wrench tectonics, such as the Bocono Fault Zone in Venezuela (Schubert 1980, Figure 2.7), are frequently developed on a similar scale to that covered by a typical Ontario gold camp. In the Timmins camp, Timiskaming - type sediments are similar in lithology and scale to the sediment fill of typical pull - apart basins. In some instances, enhanced permeability in dilational areas is a preferred locus of mineralization, relative to the compressional portions of a transcurrent shear system (Sibson 1987). This may explain the localization of gold camps along regional scale deformation zones.

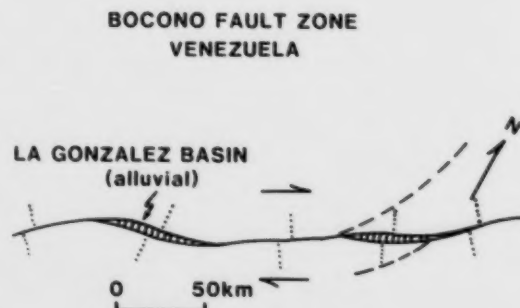


Figure 2.7. Geological sketch map of the Bocono Fault Zone, Venezuela. Note scale and shape of alluvial basins and their positioning relative to fault bends (after Schubert 1980).

2.3.2. DISTRIBUTION OF GOLD DEPOSITS

Gold deposits are composed of one or more ore zones, whose shapes, orientations and distribution can be related to small scale structures within deformation zones. Ore zones are lenticular, tabular or irregularly shaped bodies composed of veins, breccia zones, and/or stockwork systems. They are

PART I. DEPOSITIONAL MODEL

generally hosted by structures which transect stratigraphy, but some of these structures are localized along specific strata. Ore zones tend to plunge subparallel to the stretching lineation related to the shearing, as at Red Lake (Andrews *et al.* 1986), or parallel to an intersection lineation, as at Star Lake in Saskatchewan (Poulsen *et al.* 1986).

Across the Superior Province, the ore zones of the lode gold deposits are localized by:

(1) *zones of anomalously high strain within a deformation zone.* Examples include the mines of the Red Lake and Hemlo camps and the Detour Lake and Macassa Mines in Ontario, and the Sigma Mine at Val d'Or in Quebec. Both brittle (fracture and breccia vein systems) and ductile (replacement vein systems) deformation styles are recorded in these deposits, perhaps reflective of the depth in the crust at which they formed.

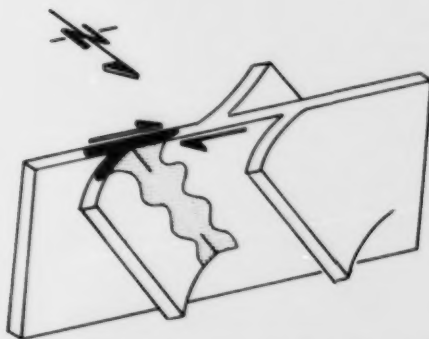


Figure 2.8. Schematic representation of ore zone control by bedding controlled splays at the Cameron Lake deposit (after Melling 1986).

(2) *pre-existing structural anisotropies.* An excellent example of this control of mineralization is the Cameron Lake deposit near Kenora (Melling *et al.* 1986), in which the ore zone occurs where sympathetic, bedding-controlled splays to a shear zone intersect that shear (Figure 2.8). The plunge of the ore zone parallels the lineation formed by the intersection of the shears. Ore zones at the Con Mine, Yellowknife (Figure 2.9), are controlled by the intersection of two shear zones (Boyle 1979).

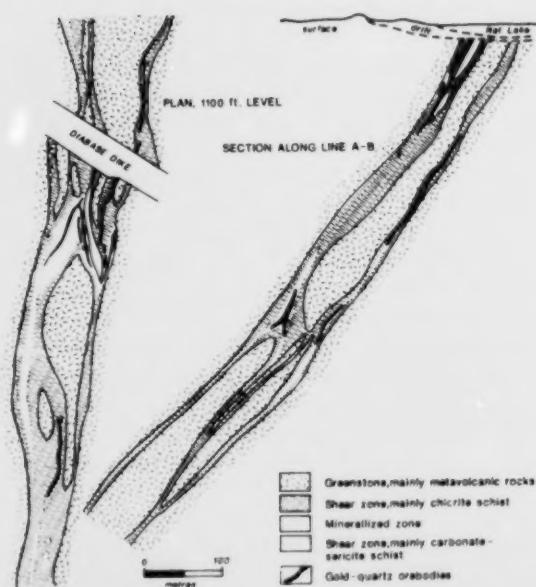


Figure 2.9. Conjugate shear systems; dilatant zones possibly developed as a result of shear zone rotation (after Boyle 1979).

(3) *a preferred lithology, where a strong competency contrast exists between adjacent rock types.* Structurally more competent lithologies can be preferentially mineralized. Examples include ore zones in the Macassa and Sigma Mines and the Dupont deposit on Lake of the Woods, in which felsic intrusive rocks contain more mineralization than surrounding, less competent lithologies. The competency difference may be a result of original lithological differences, as in the examples cited above, or may result from alteration processes. The latter control is displayed in the Dome Mine at Timmins (Hutchinson and Burlington 1984, Roberts 1987) and the Cochenour-Willans Mine at Red Lake (Sanborn-Barrie 1987), where metasomatic ankerite-rich units were more competent than enclosing lithologies and deformed in a brittle manner.

(4) *fold limbs and fold noses.* Folding of sequences of rocks of contrasting thicknesses and competencies has long been known to create permeable zones that may host mineralization. Other folding-related surfaces (eg. between layers in fold limbs, or an axial planar cleavage) may also be preferred sites for mineralization. "Saddle reef" gold deposits in fold noses have long been

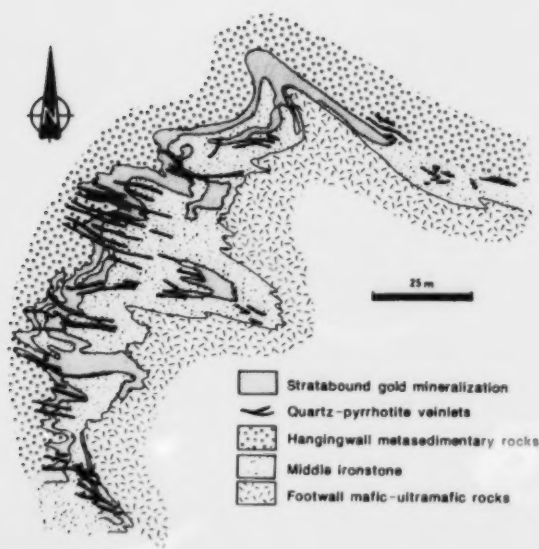


Figure 2.10. Stratabound ore occurs along fold nose; quartz-pyrrhotite veinlet mineralization is axial planar (after Hall and Rigg 1986).

known. Examples of this control of mineralization in Ontario include ore zones at Geraldton (Colvine *et al.* 1984) and the Musselwhite deposit at Opapimisk Lake (Hall and Rigg 1986), where gold is hosted in fold noses and foliation - parallel veins (Figure 2.10).

Thus specific sites or structural features within and along deformation zones may be more favourable for gold concentration.

2.3.3. DISTRIBUTION OF ORE VEINS

Many of the lode gold deposits contain an apparently complex assemblage of gold - bearing veins. However, the types of veins, their orientations, and the temporal relationships that they display can be explained in terms of the deformation processes under which they formed.

Gold - bearing veins include replacement, extension, and breccia and fracture types. In most

STRUCTURE TYPE	VEIN TYPE	NOTES
Shear Fracture	Central shear vein	Dilation of C shear
	Oblique shear vein	Dilation of P shear
Extension Fracture	Large extension veins in domains between shears	
	En-echelon gash vein arrays	May form in initial stage of shear zone development.
	Ladder vein arrays	In competent unit in ductile matrix
Extension + Shear Fracture	'Leather jacket' veins	Shear veins with envelope of gash extension veins
	'Centipede' veins	Extension-oblique shear vein combination.
Fold-Related Zones	Saddle reef	Kinematically equivalent to vein-type zone at shear intersection
	Leg reef	Shear vein in bedding on fold limb.
	Neck	Shear vein in limb thrust crossing bedding
	Spur	Extension vein normal to axial plane cleavage.

Table 2.1. Classification of shear related structurally controlled mineralized zones (after Hodgson, in press).

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CRITERIA FOR CLASSIFICATION	MAJOR CLASSES Subclasses Sub-subclasses	SPECIAL TYPES	PROBABLE GENERATING PROCESSES
Mineral or Rock-type Distribution	Vein-parallel layering Crustiform layering Single vein-wall crust Multiple stacked crusts	Phantom veinlets	Open space fill
			Open space fill of late cracks
			Crack-seal
			Decreasing width of scepta
	Lamination Sheeted veinlet zone Book texture Ribbon texture	Crinkly banding	Pressure solution
	Vein-normal layering Ladder veinlets Breccia Massive matrix Crustiform matrix Massive		Open space fill Shattering followed by open space fill Various
		Mineral Orientation	Fibrous Syntaxial Antitaxial Stretch crystal Foliated Cockade Massive
Mylonitization Open space fill Various			
Ratio of Introduced/Original Rock Minerals	Disseminated Streaky disseminated Pervasive replacement (Replacement vein)		Increasing extent of replacement of original rock
		Ghost breccia Ghost banding	Replacement of wall rock inclusions in open space fill.

Table 2.2. Textures of vein type mineralization (after Hodgson, in press).

cases, veins transect lithological contacts, and are not restricted to a specific rock type. However, veins may be statabound where they are controlled by the competency or chemistry of a particular unit. Such is the case at Geraldton, where veins occur largely in layers of iron formation (Macdonald 1984a).

In the following sections, the characteristics of each of these three types of gold-bearing vein structure are described and related to deformation processes. Much of the information in these discussions, summarized in Tables 2.1 and 2.2 and depicted in Figure 2.11, has been abstracted from a detailed review of gold-bearing veins by Hodgson (in press).

2.3.3.1. Replacement Veins

Hodgson (in press) classifies replacement veins as disseminated, streaky disseminated, and pervasive, based on increasing replacement of the original rock (Table 2.2). In general, replacement veins are highly silicified and sulphidized zones which

contain remnants of narrow, foliation-parallel quartz veinlets, quartz lenses, and quartz pods (Plate 2.3a - d). In many cases, additional deformation has caused transposition, folding or boudinage of earlier vein material (Plate 2.3e).

Replacement veins occur in ductile deformation zones and are parallel to foliation. They are generally restricted to zones of simple shear, where increased permeability results from the development of C or C' cleavages (Berthé *et al.* 1979, White *et al.* 1980, Ponce de Leon and Choukroune 1980).

Gold can be deposited in significant concentrations in this setting, either within the vein or associated with pervasive silicification, sulphidation and carbonatization of the protolith along the vein margins. Examples of deposits containing replacement veins include the Dickenson (A.W. White) Mine at Red Lake (Andrews *et al.* 1986), the Upper Canada (Roberts 1987) and Detour Lake (Marmont 1986) Mines, and in the Duport deposit (Smith 1986).

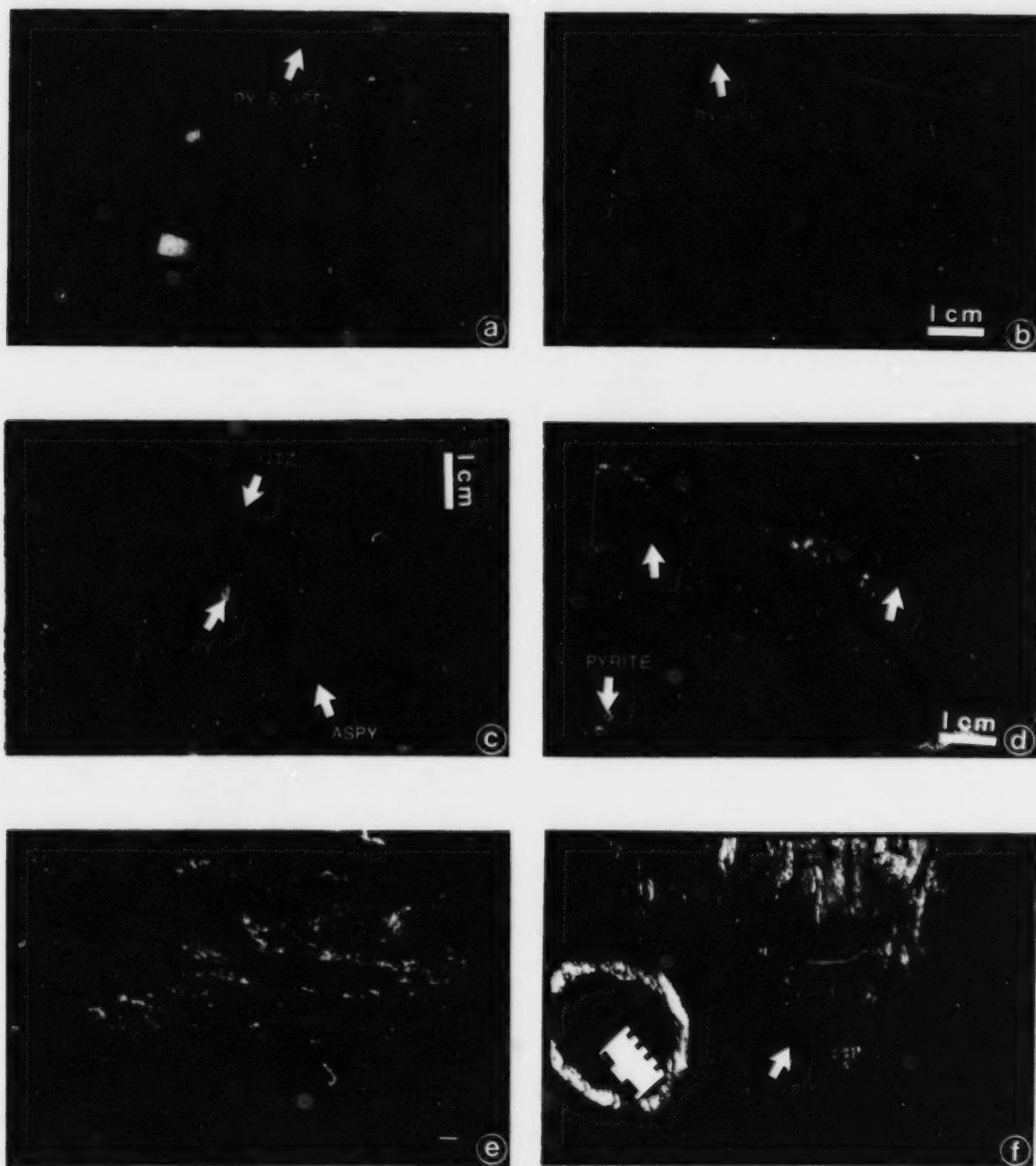


Plate 2.3. (a) Pervasive replacement ore (Dickenson (A.W.White) Mine, Red Lake).
 (b) Streaky disseminated mineralization (Duport deposit).
 (c) Foliation-parallel quartz veining in replacement ore (Duport deposit).
 (d) Transposed and boudinaged quartz vein in replacement ore (Duport deposit).
 (e) Folded quartz veins in replacement ore (Hemlo).
 (f) Replacement-type mineralization extending out into fractured and deformed pillow basalt (Dickenson Mine).

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2.3.3.2. Extension Veins

Extension veins include tension and shear varieties. Most of these veins are composite, or display crack - seal textures, with laminated or crustiform layering parallel to the vein walls (Table 2.2, Plate 2.4). These textures indicate that the veins formed as open space fillings, under oscillating, high fluid pressures (Sibson *et al.* 1975).

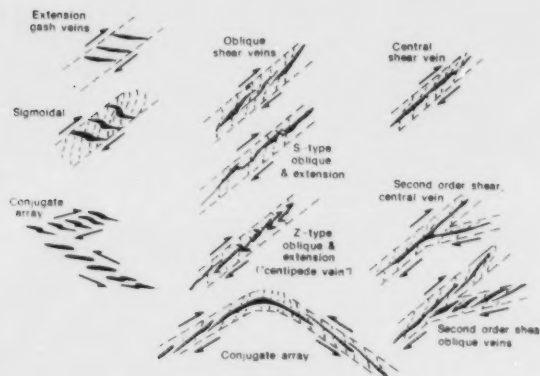


Figure 2.11. Extension vein settings (after Hodgson, *in press*).

Tension veins may develop under either pure or simple shear. They propagate parallel with, and dilate perpendicular to, the local direction of maximum principal stress (Figure 2.12). Initial mineral growth is orthogonal to the vein walls, but subsequent deformation may reopen, brecciate, rotate and fold the veins. Progressive deformation may cause new veins to develop across older, deforming ones. These veins commonly occur in *en echelon* sets, such as ladder veins which develop in more competent units. Some may have formed in the early stages of deformation, but most occur in relatively undeformed rock along the margins of deformation zones, in lithons of relatively undeformed rock within deformation zones, or between deformation zones (eg. Sigma Mine; Robert and Brown 1986).

Shear veins develop during brittle and brittle - ductile deformation under simple shear. The most common shear veins are Riedel shear fractures (Tchalenko and Ambraseys 1970). Experimental studies (Gamond 1985) indicate that R-type Riedel shears (Figure 2.13) are the most dilatant and therefore should most commonly host mineralization; however, Hodgson (*in press*) notes that most deposits are hosted by *en echelon*, oblique shear (P) and central shear (D) veins. Dilatancy may develop within Riedel and flexural shears during movement

along irregular fault planes (Newhouse 1940), due to fluid overpressuring, or along the intersection of two shears. Continued deformation may rotate, reopen, brecciate, or mylonitize shear veins.

Gold mineralization is hosted in Riedel shear fractures in many deposits: examples include the Campbell Red Lake and Dickenson (A.W. White) Mines (Andrews *et al.* 1986) and the Cochenour-Willans Mine (Sanborn and Schwerdtner 1986) at Red Lake, the Sigma Mine at Val d'Or (Robert *et al.* 1983), and the Duport deposit on Lake of the Woods (Smith 1986).

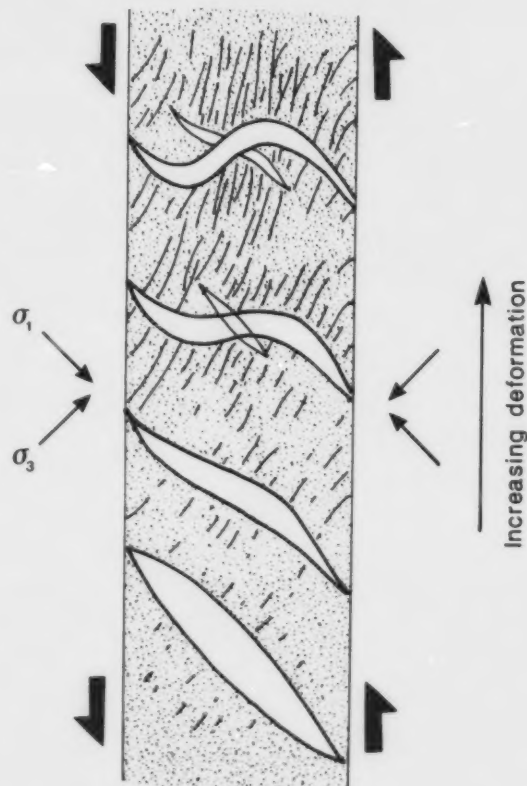


Figure 2.12. Initial development and progressive deformation of tension veins (after Kerrich and Allison 1978).

2.3.3.3. Breccia and Fracture Veins

Breccia and fracture - filling veins typically occur in the brittle - ductile domain if extremely high fluid pressures prevail. Breccias form because of frictional slip along displacement surfaces parallel with the shear zone boundary, by fracturing at fault jogs (Sibson 1987), or as a result of extremely high fluid pressures. It is the latter type that is generally associated with the gold deposits.

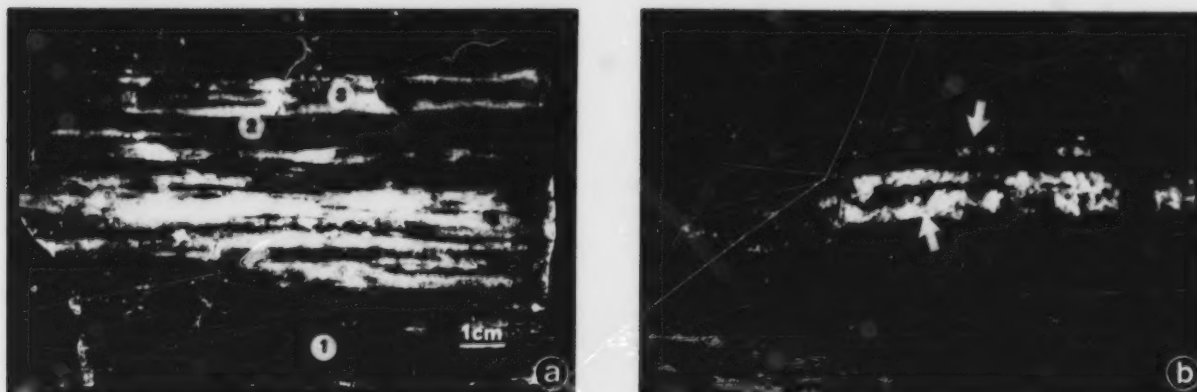


Plate 2.4. (a) Crack and seal shear vein with abundant ribbons of biotite indicating successive vein fills (Duport deposit).

(b) Crack and seal tensional vein; bands of tourmaline indicate successive vein infill (Sigma Mine).

Breccia veins typically comprise fragments of altered rock within multiple generations of matrix material. Breccia veins typically overprint earlier replacement or extension veins, but locally may be foliated, folded, or fractured during subsequent deformation. Gold occurs in the breccia fragments and matrix, and in the wall rocks adjacent to the vein. Fracture - filling veins are generally planar, and occur in regular patterns.

Examples of deposits containing breccia veins include the Macassa Mine at Kirkland Lake (Watson and Kerrich 1983) and the Duport deposit at Lake of the Woods (Smith 1986). Fracture - filling veins were mined at the Canadian Arrow and Murphy - Garrison Mines near Timmins (Cherry 1983).

Although it is a relatively simple task to classify individual gold - bearing veins in a deposit as one of replacement, extension, breccia or fracture types, it is far more difficult to unravel the complexity of vein types that typically occur in any one deposit. This complexity results from the interplay of many factors, among them the different lithologies in the deposit, the interaction of fluids with the rocks, and variations in deformation during the time that mineralization occurred.

2.4. VERTICAL ZONATION OF GOLD DEPOSITS

The ore zones and vein systems that make up gold deposits display systematic vertical and lateral changes in mineralization style. These variations reflect a change in deformation style from

brittle to brittle - ductile. For example, breccia veins occur principally within zones of brittle deformation and replacement veins typify ductile - dominated zones of deformation. Because gold deposits are both smaller than the deformation zones in which they occur and of limited vertical extent, systematic changes in the style of deformation are generally not evident within any one deposit. Whilst most deposits contain a variety of vein types, each deposit typically displays a limited number of dominant mineralization styles, which are consistent with the ambient pressures and temperatures under which that deposit formed.

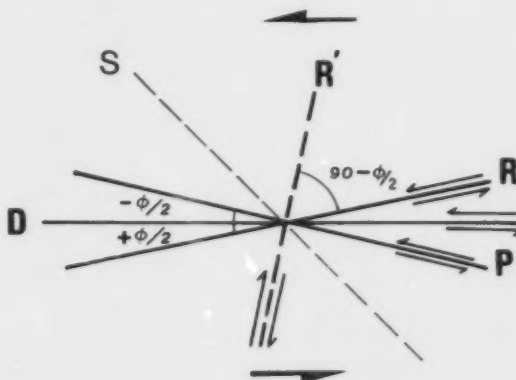


Figure 2.13. Development of Riedel shear fractures in a system of brittle simple shear (after Tchalenko 1970).

The changes anticipated in deformation zones with depth were discussed in Section 2.2.3 and illustrated in Figure 2.6 of this chapter. Figure

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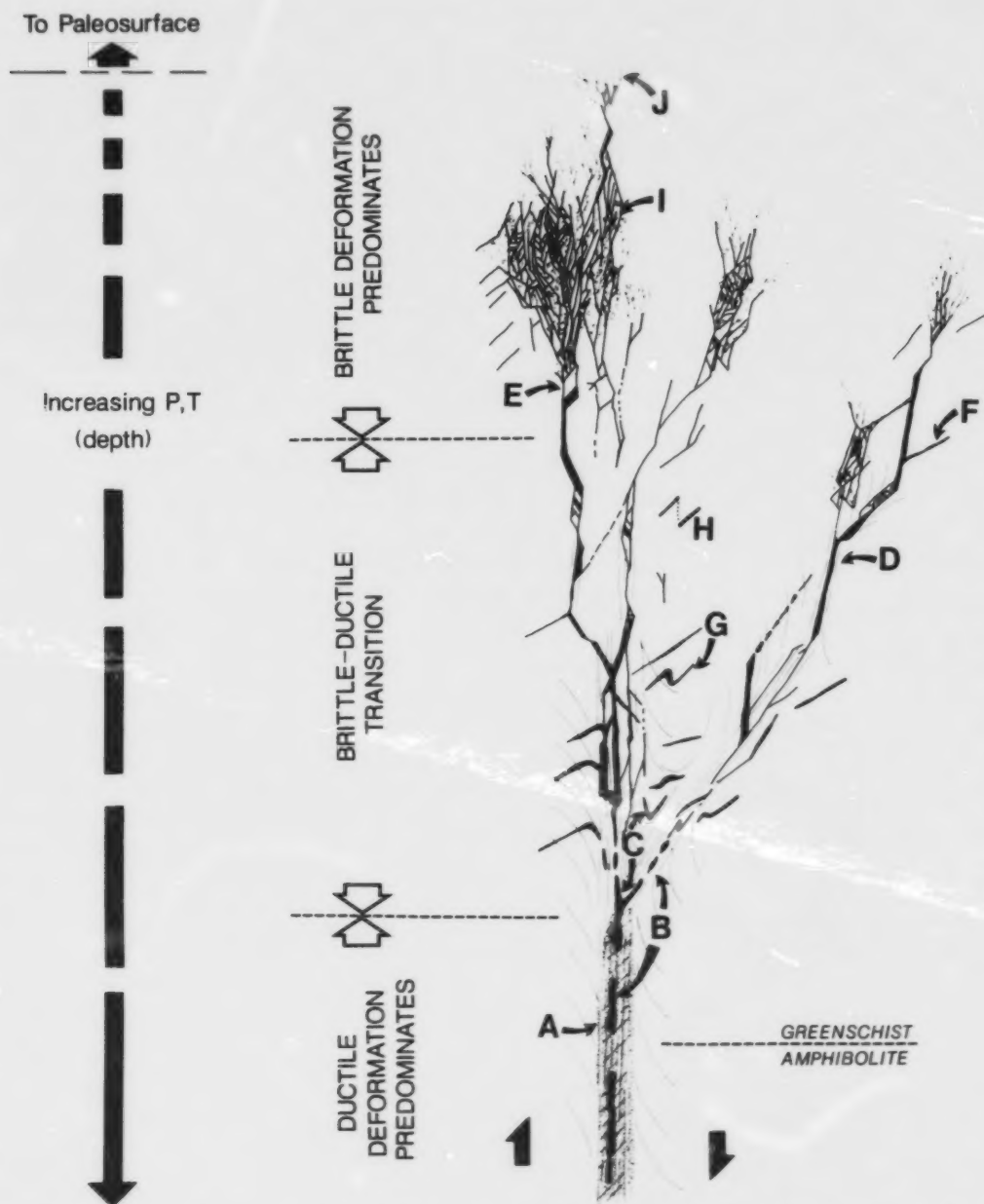


Figure 2.14. Schematic representation of an ideal ore zone. The figure illustrates: (a) replacement veins, (b) boudinaged veins, (c) saddle reef veins, (d) Riedel shear veins, (e) "break ore" veins, (f) undeformed, and (g) deformed tension veins, (i) breccia veins, (j) pervasive silicification. Vein systems are interpreted to widen and become increasingly complex with elevation. This figure does not address the effect of structural anisotropies or permutations such as changing strain rates, fluid activity or P - T variations.

2.14 illustrates idealized variations in mineralization that will occur with these changes in deformation style. This diagram is a composite, compiled from evidence summarized below:

(1) Most deposits hosted in greenschist facies rocks contain ore in shear or tension veins, or the deposits contain mainly extension, breccia and fracture - filling veins, as for example at Macassa and Canadian Arrow. In each case, the veins are products of brittle deformation. In addition, some deposits (eg. the Ross Mine near Timmins (Troop 1986)) display brittle deformation that becomes wider and more complex towards the surface. However, other deposits in greenschist terranes are hosted in ductile shear zones (e.g. Yellowknife) and contain ore shoots that are replacement veins.

(2) Deposits hosted in higher grade (upper greenschist to lower amphibolite) terranes have textures and vein types indicative of brittle - ductile deformation. Although no case has yet been documented of a deposit in rocks that have been subjected only to ductile deformation, all deposits in amphibolite facies terranes occur in rocks in which ductile deformation dominates over brittle (eg. Musselwhite (Hall and Rigg 1986), Hemlo (Hugon 1986, Kuhns *et al.* 1986), Dickenson (Andrews *et al.* 1986), Detour Lake (Marmont 1986), and Lupin (Kerswill 1986, Lhotka and Nesbitt 1987)). Most of these deposits are associated with schistose rocks and contain replacement vein systems, many of which have been folded and boudinaged. Shear veins are common, but probably reflect increased fluid pressures during deformation. Brittle structures in these deposits, such as fractures and breccias, generally reflect high fluid pressures, and occur in areas of competency contrast between adjacent lithologies, or develop later as ambient temperatures decrease.

(3) Some deposits display mineralization styles indicative of a transition in deformation style, or of a transition between different pressure - temperature regimes. For example, mineralization in the Campbell Red Lake and Dickenson (A.W. White) Mines at Red Lake exhibits vertical and lateral transitions from brittle - ductile to more brittle deformation with decreasing metamorphic grade, and decreasing depth. At the Duport deposit on Lake of the Woods, a change from brittle - ductile to brittle vein systems reflects a decrease in ambient temperature.

(4) The dominant deformation style in high grade terranes is ductile; deformation in these terranes does not appear to have produced dilatancies. Consequently, very few gold deposits occur in these terranes; those few are mainly in areas in which both chemical traps and strong competency contrasts exist. One example may be the Musselwhite deposit at Opapimisk Lake (Hall and Rigg 1986), which occurs in rocks interpreted to have been deformed at temperatures of 530° to 550°C and pressures of more than 3 kb (corresponding to a depth of more than 9 km in the crust).

Figure 2.14 illustrates the changes expected in deformation and mineralization styles with increasing depth in a single, idealized deposit, drawing from all of the evidence presented above. In the upper levels of this idealized deposit, veins would be hosted by brittle structures such as faults, tensile and shear fractures. The upper levels would also be characterized by the occurrence of hydraulic breccias, with accompanying pervasive alteration. At intermediate levels, veins would be hosted by both brittle and ductile structures, including discrete tension and shear veins and local brecciation, but also local replacement veins. Finally, at the lowest levels in the deposit, mineralization would occur in replacement and shear veins within schists. Mineralization would only occur at great depths where significant structural or chemical anisotropies generated permeability.

2.5. SUMMARY

Archean lode gold deposits in the Superior Province occur in deformation zones formed in a regime of dominantly simple shear. The characteristics of the shear zones are consistent with their formation through the variable effects of north - northwesterly - directed compression. These zones either transect or form the boundaries of greenstone belts, so they occurred as a major, late Archean tectonic event.

Gold deposits are hosted in smaller scale structures within the deformation zones. Although ore bodies and, at a smaller scale, ore veins often seem randomly oriented, it is possible to explain their distributions and orientations in terms of the simple shear processes under which they were developed. Accordingly, these structures, just as the deformation zones in which they occur, reflect the depth of burial beneath the paleosurface during deformation. With this in mind, strain analysis can

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be applied to help predict the shapes of orebodies and the styles of mineralization at any given location.

3. SYSTEMATICS OF ROCK ALTERATION

3.1. INTRODUCTION

Intense and extensive alteration is a highly visible characteristic of most Archean lode gold deposits. Alteration characteristics of these deposits vary widely, in contrast with the consistent alteration patterns associated with Archean volcanogenic massive sulphide deposits (eg. Franklin *et al.* 1981). An appreciation of the processes responsible for alteration in gold-bearing hydrothermal systems, and of the patterns produced by these processes, has both practical and theoretical value. Two important questions are addressed in this research:

- (1) Can patterns of alteration be used in exploration to target gold mineralization within more extensive chemical or mineralogical anomalies?
- (2) What can the observed alteration patterns tell us about the chemical and physical conditions prevailing during ore formation?

In this chapter, observations on alteration associated with many Archean lode gold deposits, primarily in the Superior Province, are discussed in the context of processes that might be responsible for them. In this way, it is possible to present a model to explain the observed variability in the alteration.

3.1.1. ALTERATION PROCESSES

Alteration may be defined as a change in mineralogy, chemistry or texture of rocks to more stable conditions in response to fluxes in fluid temperature, pressure and composition (Rose and Burt 1979). As such, hydrothermal alteration may be considered as a type of metamorphism, in which fluid movement, resulting largely from thermal and hydrostatic pressure gradients, is the most important medium for the transport of matter (Bickle and MacKenzie 1987, Helgeson and Lichtner 1987). As they approach equilibrium, the compositions of fluid and rock are buffered by each other (Greenwood 1975, Rose and Burt 1979). The final rock composition depends largely upon the degree of fluid:rock interaction. Metasomatism, a form of alteration, is the replacement of one mineral by another of different composition through the *in situ* addition and/or removal of components resulting from a directed interaction of hydrothermal fluid with rock. One good example of metasomatism is the hydration or carbonatization of rocks, ie. the addition of H₂O and CO₂ (Rumble 1982).

Hydrothermal fluids of magmatic or metamorphic origin are by definition H₂O-rich, although dissolved CO₂ may be present in significant amounts (Wyllie 1979, Ferry 1987). Irrespective of the composition or source of these fluids, their interaction with rock can produce profound alteration effects. This alteration is manifest principally by partial to complete mineral breakdowns, often accompanied by the addition or removal of matter in ionic solutions. Acidic fluids, for example, can effectively cause the breakdown of minerals (Anderson and Burnham 1983), as in the hydrolysis of alkali feldspar to sericite.

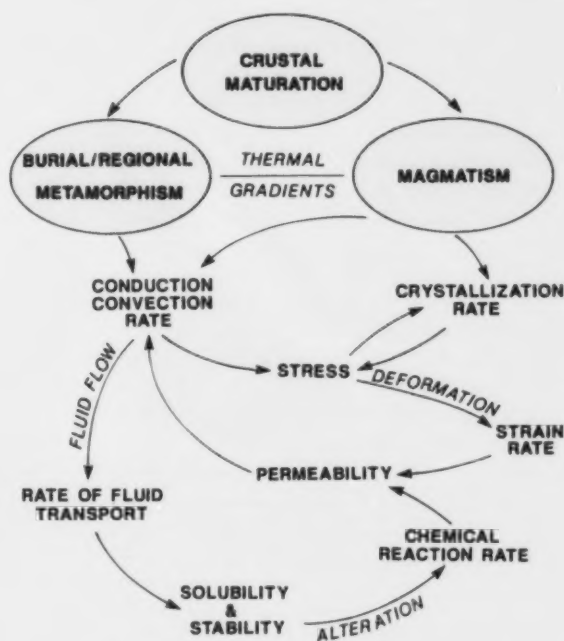


Figure 3.1. Diagrammatic representation of the interdependency between alteration and deformation and the driving processes involved (modified after Norton 1984).

Interdependent variables governing the generation of permeability and the rate and focussing of fluid flow and mass transfer in both metamorphic (Walther and Orville 1982, Etheridge *et al.* 1983, 1984, Thompson 1987) and magmatic environments (Norton and Knapp 1977, Norton and Knight 1977, Norton 1978, 1984) are summarized in Figure 3.1. The generation of pathways for fluid access during deformation, discussed in Chapter 2 of this report, is fundamental to alteration, as it controls both the rate and volume of fluid transport, and, hence, controls the amount of fluid passing through the site of

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alteration. This secondary permeability is an important control on the dimensions and intensity of alteration in hydrothermal environments (Cox *et al.* 1986).

3.2. INHERENT CHARACTERISTICS OF ALTERATION

This section addresses the gross characteristics of alteration in gold deposits. While not all of these occur in every deposit, they are frequent enough to form a core of common observations. A discussion of processes follows in the subsequent section.

3.2.1. REGIONAL DISTRIBUTION

Zones of extensive rock alteration are sporadic within the deformation zones which host the Archean lode gold deposits. Examples of deformation zones within which regional alteration occurs include the Destor - Porcupine and Kirkland Lake - Larder Lake breaks in the Abitibi belt (Bain 1933, Fyon and Crocket 1982, Hodgson 1983a, Hodgson 1986, Whittaker 1986a, b), the Mishibishu Lake (Heather 1986) and Goudreau - Lochalsh (Heather and Arias 1987) deformation zones in the Wawa belt, the Pipestone - Cameron Lake deformation zone (Buck 1986, Melling *et al.* 1986) in the western Wabigoon belt, zones in the Uchi - Confederation Lakes area (Fyon and Lane 1985, Fyon and O'Donnell 1986), and deformation zones in the Red Lake area (Durocher 1983, Andrews *et al.* 1986) of the Uchi belt.

Carbonatization (manifest as replacement by magnesite, ankerite - dolomite and calcite), restricted to these regional deformation zones, is the most prominent and common gold-related alteration on both regional and mine scales. Mineral assemblages indicative of sulphidation, silicification, oxidation, and potassic metasomatism are also characteristic, but tend to be more locally associated with gold. This carbonate alteration is distinct in its mineralogy and areal restriction from "sea floor" diagenesis and low grade metamorphism, which are dominated by calcite, epidote, actinolite and zeolites (Vallance 1965, 1974, Cann 1969, Dimroth and Lichtblau 1979). The latter mineral assemblages may be attributed largely to the hydrolysis of plagioclase and pyroxenes (eg. Dimroth and Lichtblau 1979, Jolly 1980, Gelin *et al.* 1982, Gelin *et al.* 1983, Trudel 1983, Dimroth *et al.* 1983, Walsh 1984, Barley and Groves 1987).

Deformation and alteration associated with gold mineralization can persist into at least the marginal phases of the large granitoid complexes and external gneissic terranes. Examples occur at the Renabie Mine near Wawa (Callan and Spooner 1987, Studemeister and Kilias 1987), the Flavrian and Bourlamaque batholiths near Noranda (Darling *et al.* 1986), and the internal granitic stocks in the Red Lake area (Andrews *et al.* 1986).

3.2.2. LOCAL SPATIAL AND TEMPORAL RELATIONSHIPS TO GOLD

All Archean lode gold deposits have some wall rock alteration; in most cases, the intensity of this alteration increases with proximity to mineralized structures (Hodgson 1983a, 1986). The alteration zones generally are subparallel to stratigraphy, and are cross-cut only by post-ore structures. Some lithologies are more susceptible than others to a particular type of alteration, and the expression of alteration can therefore be highly variable in different lithologies.

The occurrence of alteration minerals both in mineralized veins and in the wall rocks to mineralization suggests a close temporal relationship between the alteration process and gold introduction. This relationship is well-documented for the Timmins and the Kirkland Lake areas (Bain 1933, Hurst 1935, Fryer *et al.* 1979, Robert and Brown 1984, 1986, Hodgson 1986, Piroshco and Hodgson 1988). This temporal relationship is less clear in rocks of higher metamorphic grade (eg. Hemlo, Kuhns *et al.* 1986; Opapimisk Lake, Hall and Rigg 1986; Big Bell, Australia, Chown *et al.* 1984, Phillips and De Nooy 1988), and in rock compositions other than mafic to ultramafic (eg. sediments at Pamour No.1, Timmins, Walsh *et al.* 1986, and Casa Berardi, Quebec, Pattison *et al.* 1986). The significance of possible links between alteration and mineralization is discussed in Chapter 7, Part I and in Part II of this report.

Zonation of alteration minerals in carbonate and phyllosilicate alteration assemblages is consistent in some deposits (Heather, 1986, Smith and Kesler 1985, Robert and Brown 1986, Melling *et al.* 1986, McInnes *et al.* 1986, Phillips 1986, Piroshco and Hodgson 1988, Kishida and Kerrich 1987, Neall and Phillips 1987). In other instances, however, only minor evidence of mineral zonation is evident (Macdonald 1984b, Hall and Rigg 1986, Hamilton and Hodgson 1986, Marmont 1986, Pattison *et al.* 1986, Workman 1986). In some deposits, particu-

larly those at Timmins (Fryer *et al.* 1979, Fyon *et al.* 1983a, Burrows and Spooner 1986, Troop 1986) and Red Lake (Andrews and Hugon 1985), more complex relationships between alteration and gold may be related to an episodic series of alteration events, of which gold introduction is but one component.

Across deformational gradients, a change is observed from fracture-controlled halos of alteration in regimes of predominantly brittle deformation, to more pervasive, schistosity-controlled alteration in ductile regimes (Canadian Arrow Mine, Cherry 1983; Red Lake, Lavigne and Crocket 1983; Ross Mine, Troop 1986). Fluid flow and style of alteration are greatly affected by the style of deformation (Macdonald 1983, Phillips *et al.* 1984, Colvine *et al.* 1984); thus, just as ore structures vary with depth (see Chapter 2), so does alteration.

3.2.3. MINERALOGY OF ALTERATION

Minerals common to gold-related alteration zones, in approximate order of frequency of occurrence, include:

(1) *Carbonates*: Ferroan dolomite and ankerite, with subordinate calcite, are pervasive in many gold-bearing environments. Typically, calcite is more common in less altered rocks further from gold mineralization, although these rocks still contain Fe- and Mg-rich carbonates (Timmins, Davies *et al.* 1982; Kalgoorlie, Phillips and Brown 1987). Calcite gives way to ankerite and/or ferroan dolomite exclusively in the more highly altered wall rocks immediately adjacent to mineralized vein systems (Timmins, Fyon and Crocket 1982, Piroshco and Hodgson 1988; Mishibishu Lake, Heather 1986). Because the terms ankerite, dolomite and ferroan dolomite have been variably used, descriptions of carbonate species must be carefully assessed to determine which are actually present.

(2) *Potassic phyllosilicates*: Sericite (Hollinger, Smith and Kesler 1985; Sigma, Robert and Brown 1986) and biotite (Red Lake, Andrews and Hugon 1985; Detour Lake, Marmont 1986) are common.

(3) *Alkali feldspar*: Albite (Sigma, Robert and Brown 1984; Ross Mine, Troop 1986) or potash feldspar (Kirkland Lake, Thomson

et al. 1948; Matachewan, Sinclair 1982; Hem, Kuhns *et al.* 1986) is present in some deposits.

(4) *Chlorite*: Chlorite is locally an important alteration mineral, particularly in basalts, and is often associated with calcite or dolomite (Hollinger Mine, Mason and Melnik 1986; Ross Mine, Troop 1986; Kambalda, Neall and Phillips 1987).

(5) *Iron sulphides*: Pyrite is virtually ubiquitous (Colvine *et al.* 1984). Pyrrhotite is the predominant sulphide in some deposits, and arsenopyrite occurs locally (Opapimisk Lake, Hall and Rigg 1986; Kolar Gold Field, India, Hamilton and Hodgson 1986; St. Andrew Goldfields, Malczak 1986; Lupin, Bullis 1987).

(6) *Quartz*: Silicification, as an increase in modal quartz in the altered rock or as quartz veining and flooding, is particularly common, and is often directly associated with ore and deformation (Duport deposit, Smith 1986; Ross Mine, Troop 1986; Holt-McDermott Mine, Workman 1986).

(7) *Chloritoid, staurolite, cordierite, anthophyllite, garnet and aluminosilicates* are reported in deposits located in rocks of generally higher metamorphic grade (eg. Red Lake, Mathieson and Hodgson 1984; Casa Berardi, Pattison *et al.* 1986; Duport deposit, Smith 1986).

(8) *Diopside-hedenbergite* has been reported in several mines (Kolar Gold Field, Hamilton and Hodgson 1986; Lupin Mine, Lhotka and Nesbitt 1987).

Other alteration minerals, which may be locally abundant, include *anhydrite* and *hematite* (McIntyre Mine, Burrows and Spooner 1986; Ross Mine, Troop 1986), *Ti-bearing minerals* (Sigma Mine, Robert and Brown 1984; Kambalda, Clark *et al.* 1986), *tourmaline* (Matchewan, Sinclair 1982; Sigma Mine, Robert and Brown 1984; Timmins camp, van Hees *et al.* 1986), and *scheelite* (Hollinger Mine, Wood *et al.* 1986; Lupin Mine, Lhotka and Nesbitt 1987), plus a variety of Mo, As, Sb and base metal sulphides and sulphosalts.

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Alteration often results in marked colour changes. Carbonate minerals (in particular ankerite and ferroan dolomite) and sericite impart a waxy yellow-green colour to the altered rock on fresh surfaces, but oxidized surface exposures weather to a rusty orange colour. Ankerite - chlorite assemblages are greenish-grey on fresh surfaces (Ross Mine, Troop 1986), but also weather rusty brown. Micas containing Cr, Ba, V, and Fe often occur in association with carbonate, and impart a distinctive emerald green colouration to the rock (Kerr Addison Mine, Tihor and Crocket 1976). Biotite and hematite cause a medium brownish to reddish hue in altered rocks (Detour Lake Mine, Marmont 1986). Hematization of feldspars, which imparts a brick red colour, is common in altered granitoids and porphyries (Kirkland Lake camp, Kerrich and Watson 1984; Canadian Arrow Mine, Cherry 1983; Mishibishu Lake area, Heather 1986).

3.2.4. MINERALOGY OF GOLD

Gold concentration may itself be considered a product of the alteration process. Minerals of a suite of elements, including B, W, Sb, Te, Mo and As, commonly accompany gold, along with the alteration minerals described in the preceding section. Unfortunately, documentation of the occurrence of gold and its associated mineral suite has received far less research attention than other aspects of wall rock alteration.

The most distinctive occurrence of gold is in quartz veins. In some deposits, this is the principal location of ore, although gold also occurs in the altered wall rocks (eg. the Sigma Mine, Robert and Brown 1986). In many other deposits, however, this occurrence in veins represents only a small percentage of ore: at the Hollinger Mine, for example, more than 80% of gold ore was associated with alteration sulphides in the wall rocks to quartz veins (Graton *et al.* 1933). Plate 3.1 illustrates some of the immediate associations of gold in deposits across Ontario, as examples of the variability of its occurrence.

Although studies indicate that the gold:silver ratios of gold in the Archean deposits in Ontario average approximately 10:1, there has been no systematic investigation of how this ratio varies within a deposit, or between deposits. There are indications of zonation in gold:silver ratios peripheral to the McIntyre and Hollinger Mines at Timmins, with gold:silver decreasing away from the deposit.

Similarly, the suite of B, W, Sb, Te, Mo and As minerals that are apparently directly associated with gold has not been investigated systematically in recent years. Kerrich and Hodder (1982) presented a summary of data for these minerals, and Hodgson (1983a) documented their occurrences in specific deposits in the Abitibi greenstone belt in Ontario. While Hodgson (1983a) considered the presence of these associated minerals to be a characteristic of larger gold deposits in the Abitibi, he could not determine any pattern for their presence or absence. The exception to this irregularity in occurrence may be the tellurides, which may be more common in deposits associated with late felsic intrusions, particularly those of an alkaline affinity.

The absence of systematic studies of these mineralogical associations of gold is an obvious deficiency in the documentation of gold deposits. Much of the problem in attempting correlations of this type is the lack of precise, detailed mineralogical studies completed within mine environments. Without an adequate data base, it is impossible to evaluate if such mineral associations are of fundamental significance to understanding how Archean lode gold deposits formed.

3.2.5. GEOCHEMISTRY OF ALTERATION

Most gold deposits record massive introductions of CO₂, K₂O, S, and H₂O, in addition to Au (Kerrich 1983, Groves and Phillips 1987). Increases in As, Sb, W, Mo, B, Ag, Li, Ba, Rb and Cr occur, but do not appear to be consistent for all deposits. Local Cu, Zn, and Pb enrichment is described in many deposits (Burrows and Spooner 1986, Marmont 1986, Melling *et al.* 1986, Walsh *et al.* 1986). The reduction of iron (increased Fe²⁺/Fe³⁺ ratios) outward from vein structures is indicated in several deposits (Kerrich *et al.* 1978, Fryer *et al.* 1979). Elements such as Al, Ti, Sc, V, Zr, Nb, Ni, Cr and REE do not appear to be consistently immobile during gold-related alteration (Fryer *et al.* 1979, Kerrich and Fryer 1979, Hynes 1980, Kerrich 1983, Phillips and Groves 1984, Kerrich and Watson 1984, Ludden *et al.* 1984, Anglin and Franklin 1986, Phillips 1986, Robert and Brown 1986).

Although these element introductions are common (Fyon and Crocket 1982, Guha *et al.* 1982, Perrault *et al.* 1984, Smith and Kesler 1985, Kuhns 1986, Ramos *et al.* 1987), they have not resulted in readily recognizable alteration haloes around gold deposits. This may in part be attributed to protolith heterogeneity: original compositions greatly

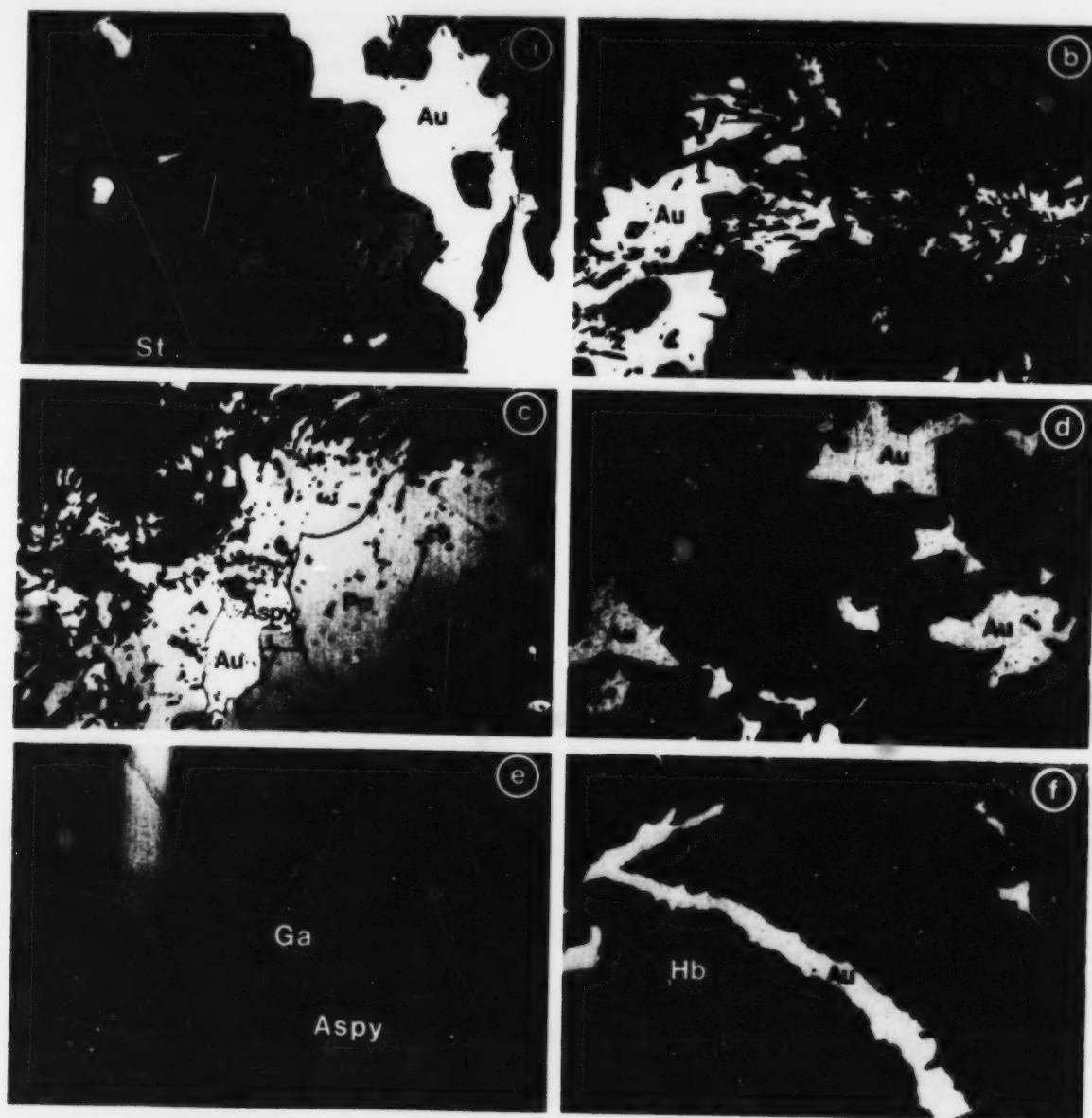


Plate 3.1. Mineralogical and textural characteristics of gold. The width of plates a, e and f is approximately 1.06 mm; the width of plates b, c and d is approximately 2.00 mm.

- A. Gold associated with pyrrhotite (po) and stibnite (Sb). From the Dickenson (A.W. White) Mine, Red Lake.
- B. Gold intergrown with fibrous amphibole. From the Dickenson (A.W. White) Mine, Red Lake.
- C. Gold and arsenopyrite (Aspy) hosted by pyrrhotite (po). From the Dickenson (A.W. White) Mine, Red Lake.
- D. Gold occurring interstitial to pyrite (py). From the Campbell Red Lake Mine, Red Lake.
- E. Gold and galena (ga) in a fracture in arsenopyrite (Aspy). From the Magnacon Mine.
- F. Cusped gold, cutting hornblende (Hb). From the Detour Lake Mine.

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influence the absolute abundances of elements in altered rocks, such that normalization is required to produce meaningful results from geochemical studies. Comparative studies of element abundances between altered and unaltered rocks require an understanding of mineral paragenesis, partition coefficients, and the activities of the components in the fluid.

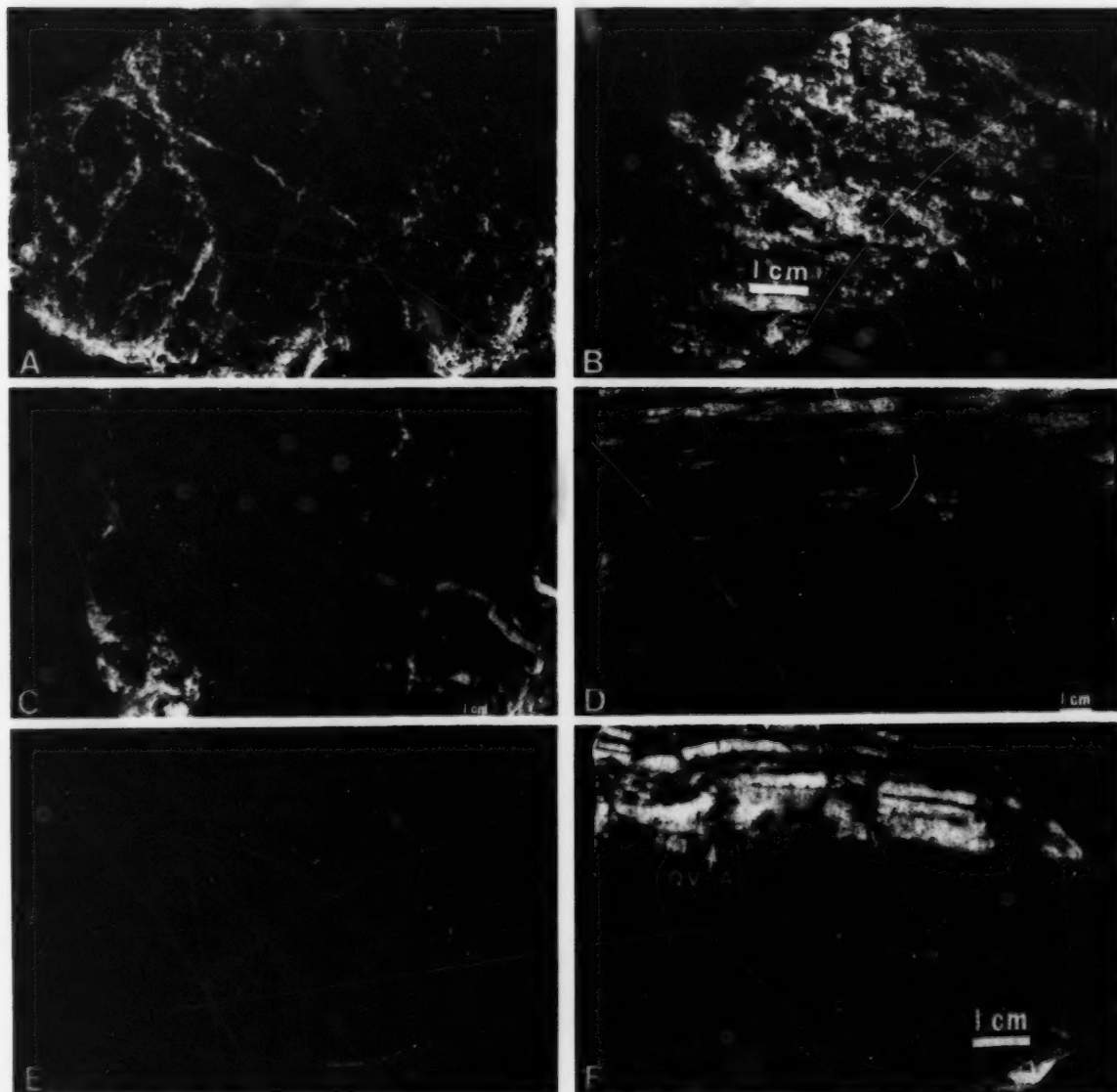
3.2.6. DIFFERENCES FROM ALTERATION IN MASSIVE SULPHIDES

Alteration associated with Archean gold deposits displays differences from that associated with volcanogenic massive sulphide deposits. These differences, which offer evidence that the compositions of the hydrothermal fluids of these two envi-

ronments were different, include:

(1) Carbonatization, which is the most characteristic alteration in Archean gold deposits, is not common in volcanogenic massive sulphide deposits (Franklin *et al.* 1981), with the notable exceptions of the Matabi (Sturgeon Lake) deposit (Franklin *et al.* 1981) and the South Bay Mine (Urabe and Scott 1983).

(2) Sericite and green mica replace chlorite and talc as the dominant phyllosilicates with increasing gold-related alteration: this is opposite to the pattern around massive sulphides (Riverin and Hodgson 1980).



(3) Alteration related to gold mineralization reflects introduction of K and CO₂ and removal of Ca and Na; alteration around massive sulphide deposits results from the introduction of Fe and Mg and the removal of Ca, Na and K.

(4) Silicification is a characteristic alteration imposed on footwall rocks associated with massive sulphide deposits (eg. Noranda, Quebec, Hall 1982, Gibson *et al.* 1983, Lesher *et al.* 1986; Matagami, MacGeehan and MacLean 1980, MacLean and Kranidiotis 1987). Andesitic basalts, which have been rendered "rhyolitic" in both composition and appearance by this silicification, may have been altered over a relatively long period of hydrothermal fluid flow. While this degree of silicification may also have occurred in gold-related alteration, it is certainly not as characteristic.

Post-alteration metamorphism of alteration assemblages can produce equally diagnostic minerals. For example, metamorphism in the footwall alteration zones of massive sulphide deposits at Noranda resulted in a cordierite - anthophyllite (dalmationite) assemblage (Spence and De Rosen-Spence 1975, Hall 1982). Similarly, staurolite, kyanite and gahnite occur at Noranda and other metamorphosed massive sulphide deposits (Snow Lake, Hutcheon 1977; Geco, Friesen *et al.* 1982). Amphibolite grade minerals in gold-related alteration zones at Red Lake include biotite, garnet, andalusite, staurolite, chloritoid, anthophyllite, kyanite and cordierite (Mathieson and Hodgson 1984). The relative timing of metamorphism and gold-related alteration is addressed in Chapter 5, Part I and in Part II of this report.

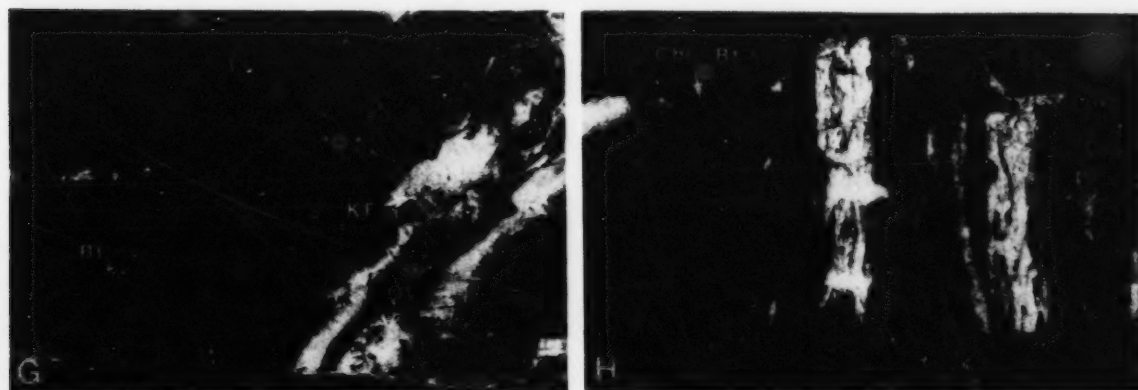


Plate 3.2. Mineralogical and textural features of alteration in komatiitic and basaltic volcanic rocks. Scale provided by one centimetre bar.

- A. Relict spinifex texture preserved by "green carbonate" fuchsite (Cr-mica) and ankerite - magnesite assemblage in altered and mineralized komatiite. From the Kerr Addison Mine.
- B. Green fuchsite (Fch) - carbonate (Cb) assemblage exhibiting banded appearance, after a probable komatiitic protolith. From the Aquarius Mine, Timmins.
- C. Intense silicification and fuchsite alteration in mineralized komatiite, exhibiting subsequent hydrofracturing and silica flooding. From the Duport deposit, Lake of the Woods.
- D. Variable sericitization (light grey areas) and carbonatization (medium grey) of a deformed variolitic basalt. From the Hollinger Mine, Timmins.
- E. Microbrecciated, quartz veined, and pervasively sericitized and carbonatized basalt. From the Ross Mine, Timmins.
- F. Altered basalt, dominated by chlorite, albite, and ferroan dolomite, exhibiting multiple fracturing and veining. Mineralized crack-seal vein (QV - Au) has no visible alteration halo. From the Ross Mine.
- G. Fractured, biotitized (Bt) basalt with intense silicification and feldspathization (KF) around mineralized quartz vein (QV). From Detour Lake Mine.
- H. Brecciated multi-stage, gold-bearing, quartz - carbonate vein (QV-Au) enclosed by biotitic alteration selvage (Bt) in pervasively altered chlorite - biotite schist (Chl-Bt). From East South C ore zone, Dickenson (A.W. White) Mine, Red Lake.

3.3. FUNDAMENTAL VARIABLES AFFECTING ALTERATION

The preceding section has reviewed the variability in alteration associated with Archean lode gold deposits. In this section, gold-related alteration is shown to be systematic when explained in terms of three fundamental variables: host rock lithology, rock permeability, and the timing and conditions of metamorphism.

3.3.1. HOST ROCK LITHOLOGY

As outlined in Chapter 1, Part I, all greenstone lithologies host gold mineralization. In this section, the control of lithologies on alteration is discussed in terms of 5 groups: mafic intrusive and extrusive rocks, komatiites, banded iron formation, pelitic sedimentary rocks and felsic intrusive rocks.

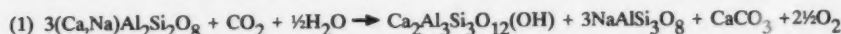
3.3.1.1. Mafic Intrusive and Extrusive Rocks

The alteration of basaltic rocks is inextricably linked to deformation: each influences both

the style and distribution of the other. For example, early pervasive fracturing and fluid flow along grain boundaries and microfractures may result in the development of a calcite + ankerite + chlorite alteration vein assemblage. Brittle fracturing and veining in this variably competent assemblage may then lead to the development of ankerite - sericite assemblages. Alteration halos to veins may extend for some distance from the most intense veining, and may be superimposed on or supersede an earlier, more widespread vein-related alteration (Robert and Brown 1986, Neall and Phillips 1987, Piroso and Hodgson 1988).

Pervasive, rather than vein- or fracture-related, alteration may occur in areas peripheral to and within the zone of gold concentration and generally within highly strained rocks (Fyon *et al.* 1982). Selective alterations of deformed volcanic textures such as pillow selvages, varioles, and hyaloclastites are common, as observed at the Hollinger Mine (Plate 3.2c, also Mason and Melnik 1986), the St. Andrew Goldfields deposit (Malczak 1986) and the Ross Mine (Troop 1986) in the Timmins - Kirkland Lake area.

Breakdown of Plagioclase Feldspar:



Plagioclase

Epidote

Albite

Calcite

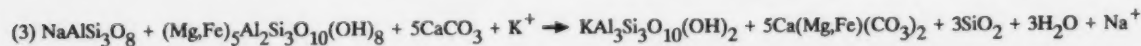
Hydrolysis of Albite:



Albite

Sericite

Quartz



Albite

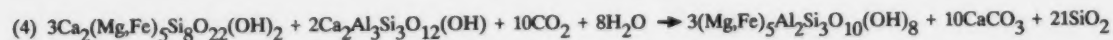
Chlorite

Calcite

Sericite

Ankerite-Dolomite Quartz

Progressive Breakdown of Hydrous Ferromagnesium Minerals:



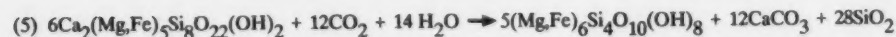
Actinolite

Epidote

Chlorite

Calcite

Quartz

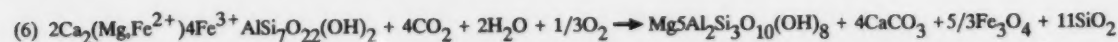


Actinolite

Chlorite

Calcite

Quartz



Amphibole

Mg - Chlorite

Calcite

Magnetite

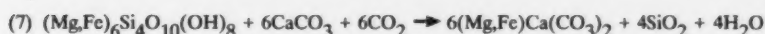
Quartz

Many gold-mineralized and altered basalts are enriched in quartz, either in discrete veins, or disseminated in the wall rocks adjacent to the veins. This increase in modal quartz is not necessarily a result of SiO_2 enrichment, since SiO_2 may remain constant or even be depleted during alteration, as a result of desilicification reactions (Table 3.1, Reactions 2 to 8). Auriferous fluids may transport large quantities of silica, but some quartz is almost certainly of local derivation. This argument is consistent with the crack-seal nature of many quartz veins in gold deposits (McKinstry and Ohle 1949), in which silica deposition may occur over an extended history of mineral reaction and vein formation.

A schematic representation of progressive mineral changes in basaltic rocks from unaltered to highly altered, indicative of increasing interaction with fluid, is shown in Figure 3.2. The fields of this and other, similar diagrams in this chapter have been positioned to reflect the common mineral associations in altered, gold-bearing rocks; they do

not represent rigorous determinations of mineral phase equilibria. Many of the possible reactions which could generate the assemblages on Figure 3.2 are given in Table 3.1. These reactions have been written to be consistent with observed geochemical mass balances for altered, gold-bearing basalts: K, H_2O , CO_2 and S have been added, and Na may have been removed. Paths of increasing alteration intensity and increasing metamorphic grade on the diagram indicate possible sequential changes in alteration mineral assemblages.

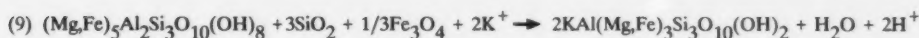
Zonation in gold-related alteration mineral assemblages is best shown in basalts by the juxtaposition of a chlorite - calcite assemblage with sericite - ferroan dolomite wall rock adjacent to veins (see Table 3.1, Reactions 3, 8). Na_2O is stripped from areas of highest fluid:rock ratio, eventually resulting in the breakdown of plagioclase feldspar (Table 3.1, Reactions 1, 2, 3). Sulphide-rich mineralized wall rocks and the breakdown of Fe-silicates in areas of mineralization indicate that sulphidation is



Chlorite Calcite Ankerite - Dolomite Quartz



Chlorite Calcite Sericite Ankerite - Dolomite Quartz



Chlorite Quartz Magnetite Biotite

Formation of Chloritoid under High XCO_2 :



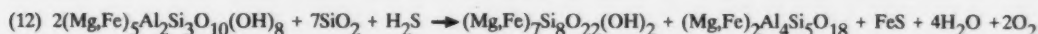
Chlorite Calcite Chloritoid Ankerite - Dolomite Quartz

Development of Rutile under High XCO_2 :



Sphene Rutile Calcite Quartz

Sulphidation of Chlorite:



Chlorite Quartz Anthophyllite Cordierite Pyrrhotite

Table 3.1. Some possible alteration and metamorphic mineral reactions in mafic rocks during gold introduction.

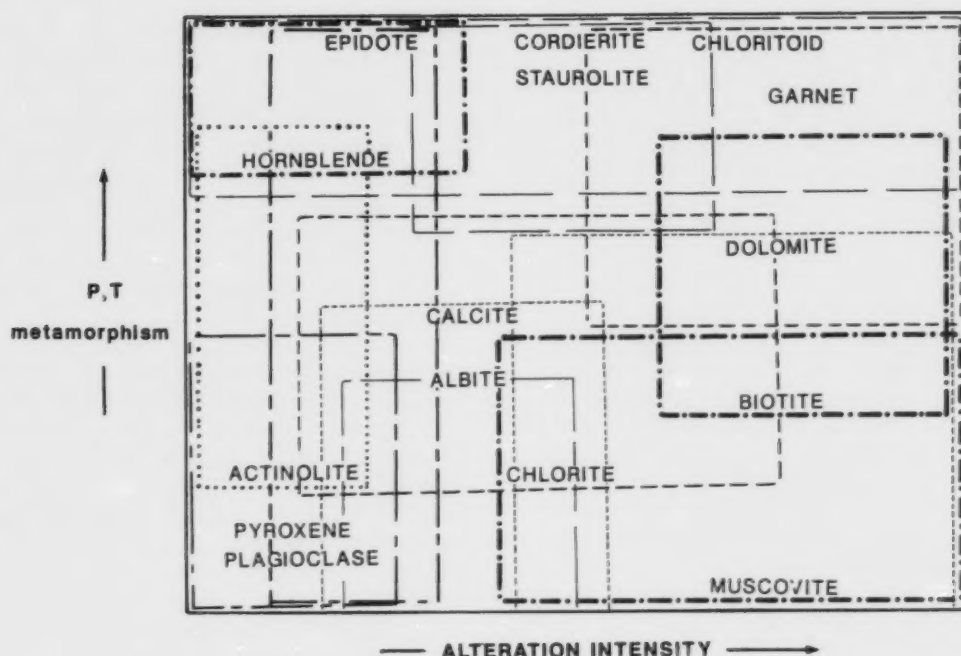


Figure 3.2. Schematic representation of the fields of occurrence of important minerals in progressively altered and metamorphosed basaltic rocks in gold-bearing systems. Note that neither metamorphism nor alteration intensity is scaled; only relative increase in grade or fluid:rock ratio is indicated.

an important process in the alteration (Table 3.1, Reaction 12). For example, the Mg/Fe + Mg ratio of chlorite increases with proximity to sulphide mineralization at the Hunt Mine, Kambalda (Neall and Phillips 1987).

Hydrolysis of feldspars is responsible for the peraluminous composition of altered basalts (Table 3.1, Reactions 1, 2). Phyllosilicate minerals such as chlorite, sericite and biotite stabilize from these compositions at low pressure - temperature conditions, and garnet, staurolite, cordierite and anthophyllite stabilize at higher metamorphic grades (Figure 3.2, Table 3.1, Reactions 2 to 6, 8 to 10, 12). However, metamorphic mineral stability fields within the basaltic system are not completely defined, largely as a result of solid solutions in the minerals (Harte and Graham 1975, Laird 1980, Apter and Liou 1983, Maruyama *et al.* 1983). In some gold deposits, textural evidence suggests that the stability ranges of some greenschist minerals have been extended to amphibolite grade temperatures (Andrews *et al.* 1986, Clark *et al.* 1986).

3.3.1.2. Komatiitic Volcanic Rocks

Komatiitic volcanic sequences are geologically and petrologically well constrained in many areas, including the Abitibi belt (Arndt 1986, Arndt *et al.* 1977, Arndt and Nesbitt 1982, Ludden and Gelinis 1982). Many of the "talc - chlorite - carbonate schists" commonly referred to in the older literature on Archean gold deposits are altered komatiites. One of the better examples of gold hosted largely within komatiite is the Kerr Addison Mine at Larder Lake (Kishida and Kerrich 1987). Ultramafic rocks are also important ore hosts at the Detour Lake Mine (Marmont 1986), and the Campbell Red Lake and Dickenson (A.W. White) Mines in Red Lake (MacGeehan and Hodgson 1982, Andrews and Hugon 1985).

The mineral assemblage typical of unaltered komatiites (olivine + clinopyroxene + lesser orthopyroxene + calcic plagioclase) is moderately well preserved in the Abitibi belt (Arndt *et al.* 1977, Canil 1987). These compositions are vulnerable to serpentinization (Wicks and Whittaker 1977, Purvis 1984) and carbonatization (Beswick 1983, Arndt and Jenner 1986) in hydrothermal environments, and also to deformation.

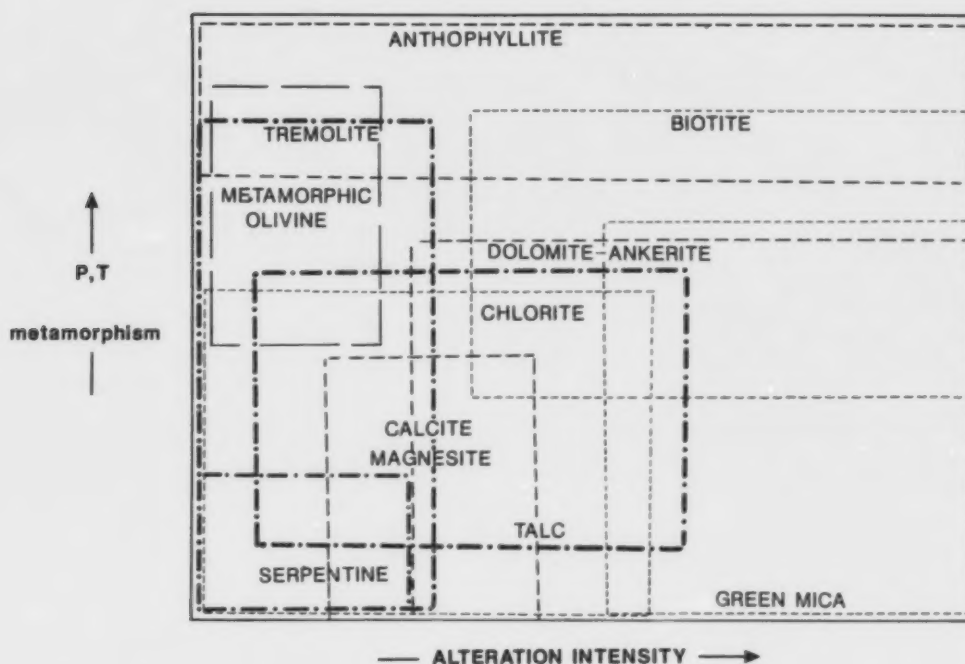


Figure 3.3. Schematic representation of the fields of occurrence of important minerals in progressively altered and metamorphosed ultramafic and komatiitic rocks in gold-bearing systems. Scaling is the same as in Figure 3.2.

The mineralogy of altered komatiites is generally simple and can be directly related to the dilution of a specific protolith composition during reaction with hydrothermal fluid. Schematic mineral stabilities in komatiites as a function of fluid:rock ratio and metamorphic grade are portrayed in Figure 3.3. Characteristic to gold-related alteration in ultramafic rocks are "green carbonate" assemblages of green mica + ankerite-dolomite (Table 3.2, Reactions 3, 4, 5; see also Malczak 1986, Kishida and Kerrich 1987), as well as tremolite, talc, biotite, and anthophyllite (Table 3.2, Reactions 1, 2, 6). Relict spinifex texture may be pseudomorphed by green mica in highly carbonatized komatiites, indicating only minor wall rock volume change during alteration (Plate 3.2a, b). Potassic metasomatism results in the formation of hydrothermal biotite at higher metamorphic grades, producing brownish halos about mineralization (Andrews and Hugon 1985, Smith 1986).

The phyllosilicate-rich mineralogy developed in altered komatiites favours a generally strong ductile response during deformation (Plate 3.2b), and mineralization is often localized in the contacts between ultramafic schists and the more brittle deformed basalts (Andrews and Hugon 1985, Marmont 1986). An exception to this ductile behaviour

is the Duport deposit (Smith 1986), where brittle hydrofracturing and silica flooding followed silicification, rather than carbonatization, in mineralized fuchsite komatiite (Plate 3.2c). Highly deformed and mineralized "flows" are common in komatiite-hosted gold deposits (Kishida and Kerrich 1987).

Geochemical studies of alteration in komatiites may produce inconsistent and misleading results (Kerrich 1983). This can primarily be related to the chemical effect that crystal fractionation and flow differentiation have on the chemistry of individual komatiitic flows (Arndt and Jenner 1986), onto which later hydrothermal alteration may be superimposed. Normalization to separate the chemical variation due to crystal fractionation from that related to gold alteration is particularly pertinent to quantifying alteration in komatiites (Beswick 1983, Kishida and Kerrich 1987).

3.3.1.3. Banded Iron Formations

The association of banded iron formations (BIF) with gold ore is common, and has provided some impetus for the syngenetic, exhalative origin for Archean gold deposits (Fripp 1976, Fryer *et al.* 1979), as most BIF's are commonly believed to have formed by chemical sedimentation (Gross 1980).

PART I. DEPOSITIONAL MODEL

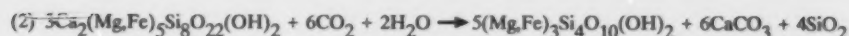
Breakdown of Hydrous Mg - Fe Silicates under High XCO_2 :



Serpentine

Talc

Magnesite

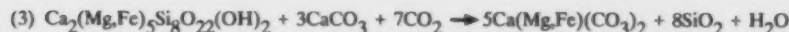


Tremolite

Talc

Calcite

Quartz

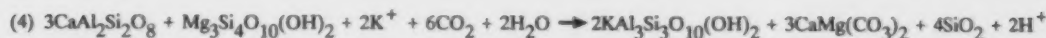


Tremolite

Calcite

Dolomite - Ankerite

Quartz



Anorthite

Talc

Green Mica

Dolomite

Quartz

Progressive Carbonation and Dehydration:



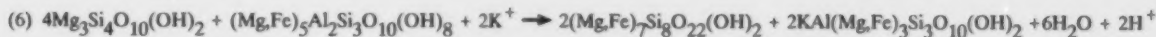
Talc

Calcite

Dolomite

Quartz

Development of Anthophyllite and Biotite at Higher Grades:



Talc

Chlorite

Anthophyllite

Biotite

Table 3.2. Some possible alteration and metamorphic mineral reactions in ultramafic rocks during gold introduction.

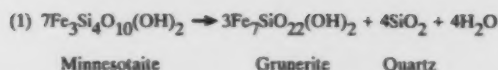
Gold mineralization is often localized within one horizon of BIF. However, many studies have demonstrated that a high degree of systematic structural control exists in BIF-hosted lode gold ores (Lavigne and Crocket 1983, Macdonald 1983a, Phillips *et al.* 1984, Groves *et al.* 1987) and that sulphidation reactions are critical to controlling the gold and sulphide content of the BIF (Fyon *et al.* 1983, Macdonald 1984a, Phillips *et al.* 1984, Macdonald and Fyon 1986).

Both the alteration and metamorphism of iron formation under varying pressure - temperature - fluid conditions are well constrained by field and experimental studies of mineral assemblages (Floran and Papike 1978, Klein 1978, Frost 1979, Gole and Klein 1981, Klein and Gole 1981, Haase 1982). Thus, BIF offers a good opportunity to compare mineral assemblages related to progressive metamorphism with those caused by gold-related alteration. The schematic stability fields of these minerals are shown on Figure 3.4.

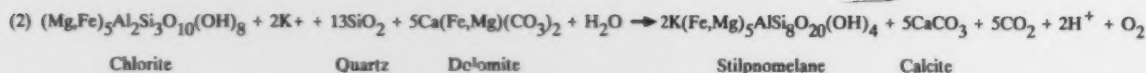
The silicate mineralogy of BIF's shown on Figure 3.4 can be subdivided into three broad groups: primary, metamorphic and hydrothermal. Primary (or diagenetic) iron silicates such as greenalite and minnesotaite (which would be converted to grunerite under prograde metamorphism, Table 3.3, Reactions 1, 3) are altered to stilpnomelane and dolomite adjacent to veins in the chert - pyrite BIF at the Agnico Eagle Mine, Quebec (Wyman *et al.* 1986). Thus, in some cases, minerals such as stilpnomelane (Table 3.3, Reaction 2), which have previously been identified as prograde metamorphic minerals in gold deposits, may have formed as a direct result of interaction between BIF and a gold-bearing fluid.

The alteration of specific BIF layers to a more peraluminous and potassic composition has resulted in the formation of biotite and garnet (Table 3.3, Reaction 8) at Opapimisk Lake (Hall and Rigg 1986), of muscovite and garnet at the Bousquet deposit (Valliant and Barnett 1982), and of stilpnomelane and hornblende (Table 3.3, Reac-

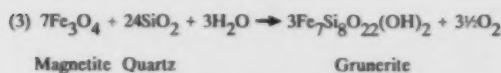
Dehydration of Minnesotaitite:



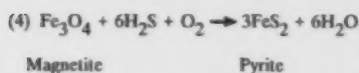
Formation of Stilpnomelane:



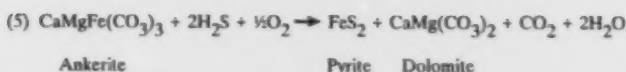
Reduction of Magnetite:



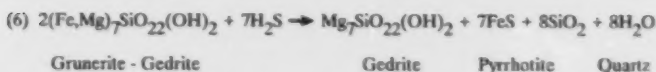
Sulphidation of Magnetite:



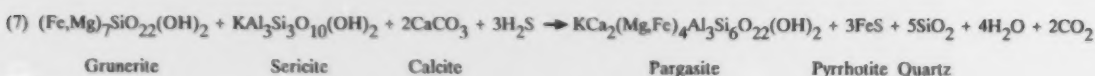
Sulphidation of Iron Carbonate:



Sulphidation of Grunerite:



Sulphidation of Grunerite to Hornblende:



Production of Garnet in BIF Interbeds:

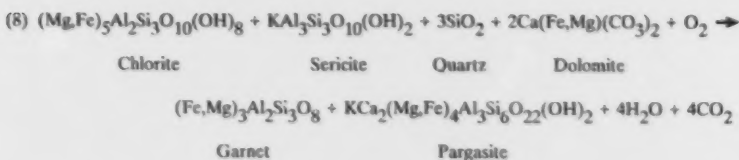


Table 3.3. Some possible alteration and metamorphic mineral reactions in oxide-silicate iron formation during gold introduction.

tion 2, 7) at the Lupin Mine (Bullis 1987, Lhotka and Nesbitt 1987). Pervasive iron carbonatization of silicate laminae within an oxide facies BIF occurs at the MacLeod-Cockshutt Mine in Geraldton (Macdonald 1983a), but carbonate replacement is not commonly observed in gold-altered BIF.

Sulphidation in BIF (Table 3.3, Reactions 4 to 7) occurs principally as discrete halos to and within fractures, or as "roll fronts" along specific compositional layers. Sulphidation results in one or more of pyrite, pyrrhotite, and arsenopyrite replacing iron oxides (Plate 3.3a, b, c; also Fyon *et al.* 1983,

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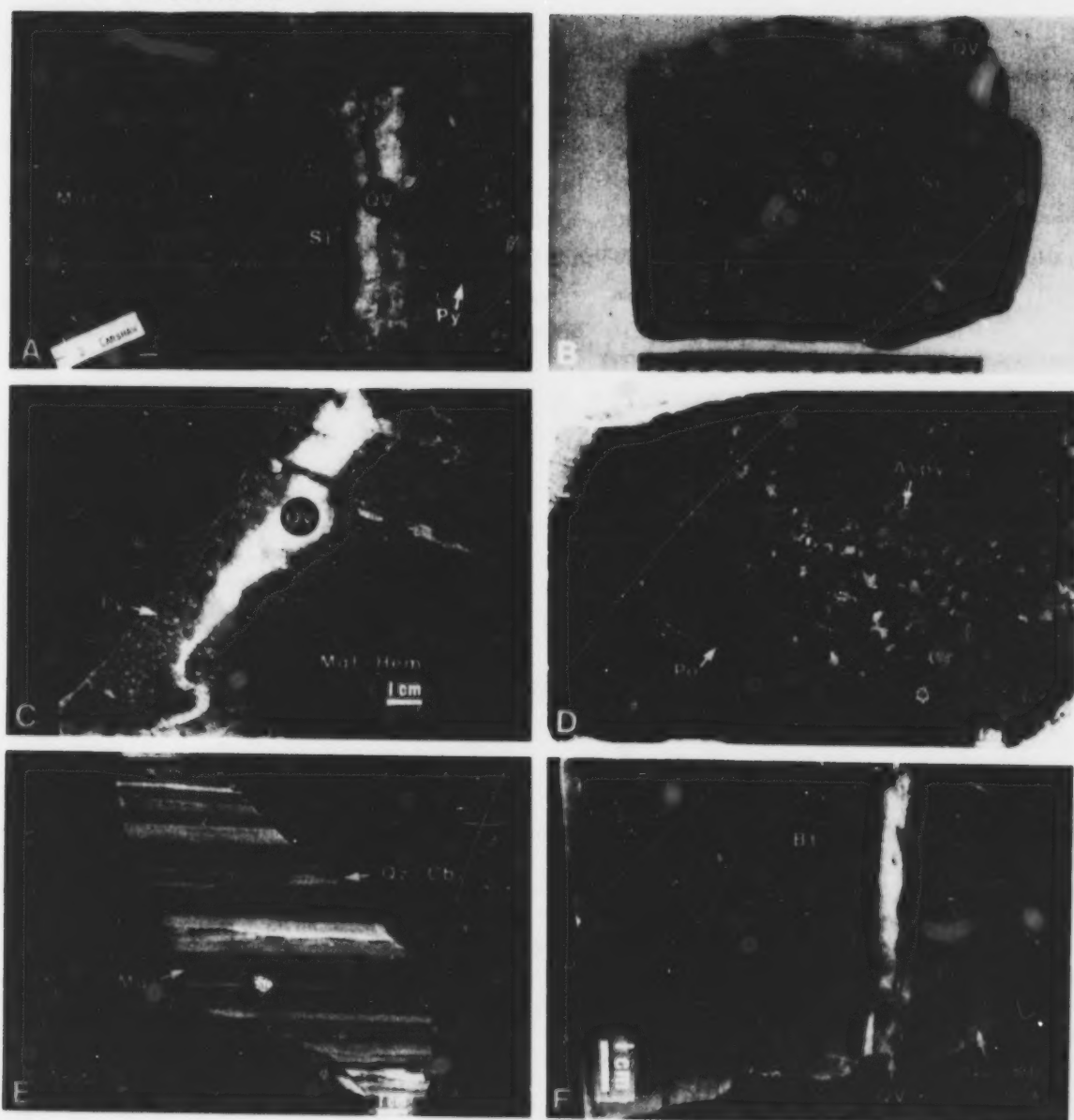


Plate 3.3. Mineralogy and textures of alteration in banded iron formations. One centimeter scale bar shown.

- A. Halo of silicification (Si) and pyritization (Py) about a massive quartz vein (QV) in banded magnetite (Mgt) iron formation. From Carshaw deposit, Timmins.
- B. Extensive silicification (Si) and pyritization (Py) related to quartz veining in magnetite (Mgt) iron formation. Note the fracture-controlled encroachment of pyrite perpendicular to the replacement bands. From Carshaw deposit, Timmins.
- C. Complete sulphidation halo of pyrite (Py) around a quartz vein (QV) in banded magnetite - hematite (Mgt-Hem) iron formation. From the Solomon's Pillars deposit, Geraldton.
- D. Bands of sulphidation by pyrrhotite (Po) in a silicate (grunerite) iron formation, with superimposed growth of arsenopyrite during fracturing (?). From Lupin Mine, Northwest Territories.
- E. Pervasive carbonatization of sedimentary interbeds in banded magnetite iron formation. Altered interbeds weather a rusty colour due to ankerite content. From Hard Rock Mine, Geraldton.
- F. Highly silicified, feldspathized (Qz-Kf) and biotitized (Bt) basalt which could be misidentified as a banded chert iron formation. Compare with 2E. Note the alteration halos to cross-cutting fractures. From the Detour Lake Mine.

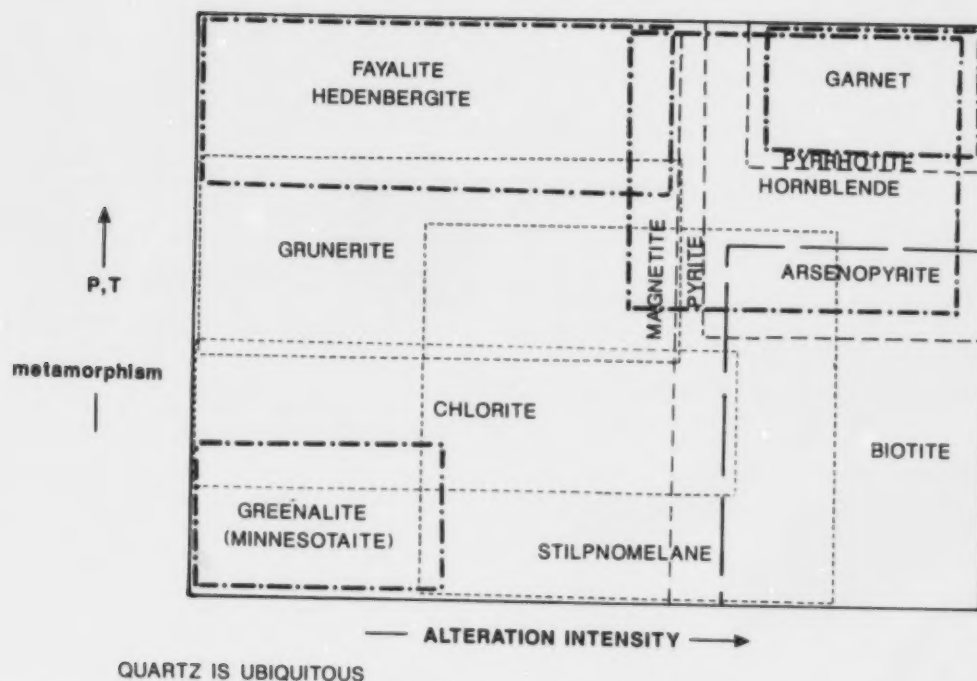


Figure 3.4. Schematic representation of the fields of occurrence of important minerals in progressively altered and metamorphosed silicate-oxide facies iron formation in gold-bearing systems. Scaling is the same as in Figure 3.2.

Wyman *et al.* 1986, Bullis 1987). At the Lupin Mine, Lhotka and Nesbitt (1987) demonstrate a systematic decrease in the Fe/Fe + Mg ratio of amphiboles and an increase in total sulphide content within the mineralized BIF, indicating that sulphidation is an important process in controlling both mineral assemblage and composition (Table 3.3, Reactions 6, 7).

3.3.1.4. Pelitic Sedimentary Rocks

Few Archean deposits are entirely hosted within clastic sedimentary rocks, but many do contain them, as at Hemlo (Muir 1983, Burk *et al.* 1986), Geraldton (Macdonald 1983a), Timmins (Walsh *et al.* 1986), and Contwoyto Lake (Kerswill 1986).

Pelitic sedimentary rocks commonly respond to shear in a ductile fashion, and hydrothermal fluids may be preferentially focussed through adjacent, more competent rocks such as basalts or BIF, which are also geochemically more favourable for sulphidation and mineralization. Well-developed vein

systems are observed to terminate abruptly against argillaceous units, illustrating the effects of lithologic competency contrasts during deformation and fluid migration (Downes *et al.* 1984, Pattison *et al.* 1986). Thus, the physical characteristics of some sedimentary rocks may deter fluid flow and these rocks would be less favourable for gold-related alteration and mineralization.

Particularly common in gold-related alteration of sedimentary rocks is the formation of micas and K-feldspar, representative of introduction of K₂O. This was demonstrated for the Pamour No. 1 Mine at Timmins by Lickley *et al.* (1987), who devised a method of normalizing the composition of mineralized sediments by separating the arenaceous (feldspar-rich) and argillaceous (mica-rich) components along a mixing curve, and thereby quantifying significant K₂O anomalies in mineralized sedimentary wall rocks.

Other studies (Macdonald 1983a) have identified elevated carbonate and sulphide mineral contents in sedimentary rocks within some gold deposits, but these extend only a few metres from alteration-related veins.

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3.3.1.5. Granitoids and Felsic Rocks

Felsic rocks behave competently during deformation, and therefore commonly host fracture-related mineralization and alteration (Plate 3.4e, f). Excellent examples of restricted fracture-halo alteration in granitoids occur at the Canadian Arrow Mine in the Kirkland Lake area (Cherry 1983, Plate 3.4f), at the Kirkland Lake mines (Thomson *et al.* 1948, Kerrich and Watson 1984), and at the Lamaque Mine in Quebec (Perrault *et al.* 1984). Ductile shearing is also observed in porphyries, in which cases pervasive carbonatization and sericitization are concomitant with fabric development (Plate 3.4a, b; also Davies and Luhta 1977, Burrows and Spooner 1986, Callan and Spooner 1987).

The characteristic minerals of granitoids and felsic volcanic rocks have wide stability ranges, in marked contrast to those of basalts or ultramafic rocks. Mineral assemblages in progressively altered granitoids are schematically illustrated in Figure 3.5; two applicable mineral reactions appear in Table 3.4 and others are in Tables 3.1 to 3.3.

Albite is a common alteration mineral in felsic rocks. Its formation is probably controlled by the hydrolytic breakdown of alkali feldspar to albite

plus sericite (Table 3.4; Reaction 1, Plate 3.4c, d). Albite and sericite (plus calcite and chlorite) may form from the breakdown of plagioclase and magmatic biotite (Plate 3.4h; Table 3.4, Reaction 2); one example of this may be in tonalites at the Renabie Mine (Studemeister and Kilias 1987). These reactions produce large quantities of silica, reflected by the quartz-rich groundmass of altered porphyries and the common multi-stage quartz veining (Plate 3.4g; Cherry 1983). Secondary hematite-impregnated orthoclase overprinting primary igneous feldspars has been identified in altered syenites of the Kirkland Lake - Matachewan area (Sinclair 1982, Kerrich and Watson 1984).

The spatial relationship of oxidized minerals such as anhydrite and hematite (eg. Timmins, Ross Mine, Plate 3.4c, d; Matachewan, and Kirkland Lake) with porphyry intrusions reflects the strong oxidizing effects produced during their emplacement and crystallization (Cameron and Hattori 1987). Burrows and Spooner (1986) present some evidence to suggest that the anhydrite pre-dates the gold at the McIntyre Mine, Timmins, but the relative timing of intrusion, alteration and gold introduction has not been established unequivocally.

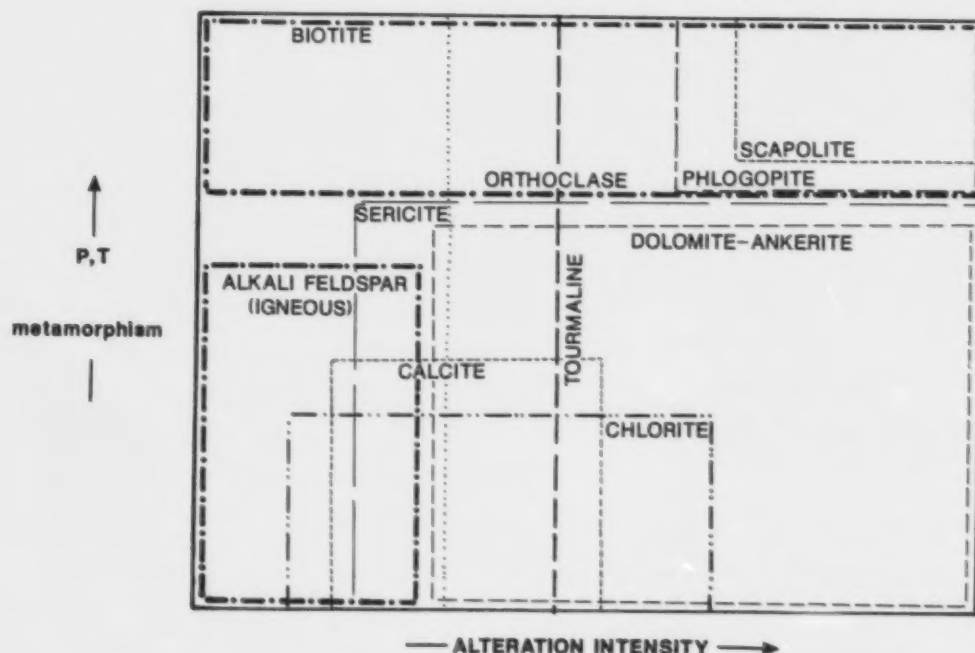
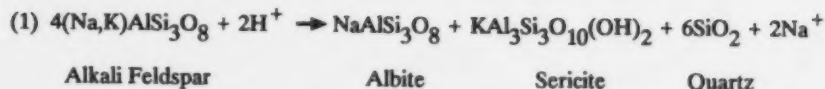


Figure 3.5. Schematic representation of the fields of occurrence of important minerals in progressively altered and metamorphosed silica-saturated and -undersaturated granitoid rocks in gold-bearing systems. Scaling is the same as in Figure 3.2.

Hydrolysis of Alkali Feldspar:



Formation of Albite - Sericite - Chlorite - Calcite Assemblage:

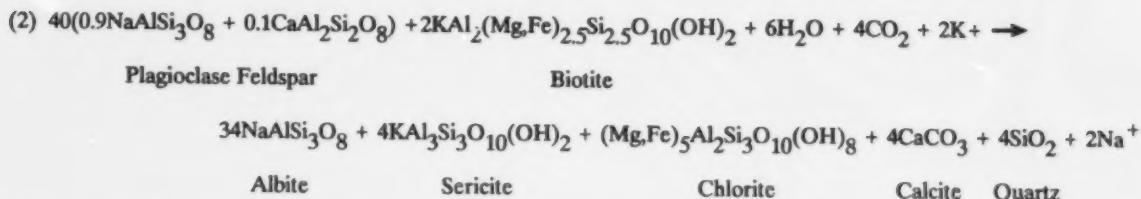


Table 3.4. Some possible alteration and metamorphic mineral reactions in granitoid rocks during gold introduction.

3.3.2. PERMEABILITY AND THE FLUID:ROCK RATIO

Rock permeability is fundamental to alteration: the hydrothermal fluid must pass through the rock to alter it. As discussed in Chapter 2, local deformation is the principal mechanism of generating enhanced permeability in alteration zones.

Permeability effects the rate and extent of fluid flow, and hence the degree of interaction between minerals and fluids. Brittle deformation typically generates faults, fractures and other open spaces through which the fluid passes. Mineralization related to these structures occurs as discrete veins or lodes, with alteration typically being more intense adjacent to the mineralized structures. Such open spaces are not common in rocks undergoing ductile deformation. In such cases, the fluid may be transported through the rock along closely spaced surfaces such as schistosity. The mineralization and alteration associated with this type of deformation are often disseminated through the rock. Complexities to this apparently simple pattern arise from changes that occur during deformation, alteration and mineralization processes. For example, as discussed in previous sections of this chapter, and in Chapter 2, early alteration of some lithologies may change their deformation style from ductile to brittle, or *vice versa*. This can result in the superposition of one mineralization and alteration style upon another.

The fluid:rock ratio represents an estimate of the volume of fluid that has passed through a given volume of rock. The alteration effects of a unit volume of fluid passing through discrete frac-

tures might be expected to differ from the effects of the same volume of fluid passing through rocks along closely spaced planar structures, as described above. In general, however, the systematic changes in mineralogy outlined in the preceding section should proceed further as the volume of fluid passing through the rocks increases. Thus, the total volume of fluid affecting the alteration also has a fundamental control on the resulting mineralogy.

3.3.3. THE ROLE OF METAMORPHISM

Evidence is presented elsewhere in this report to support the premise that gold was introduced across the Superior Province at near peak metamorphic pressure - temperature conditions in the greenstone terranes. As well, metamorphic dehydration has been proposed as the major process responsible for generation of gold-mineralizing fluids (Groves and Phillips 1987, Groves *et al.* 1987). Both suggestions require that the possible relationship between metamorphism and alteration be examined.

3.3.3.1. Pressure - Temperature Conditions and Alteration

Most Archean gold deposits occur in terranes which have a mineralogy indicative of peak metamorphism at either greenschist or amphibolite grade conditions. The mineral assemblages in pelitic sediments and mafic volcanics, and mineral geothermometers and geobarometers, provide the best constraints on the pressure - temperature - fluid conditions of metamorphism. The characteristic alteration mineral assemblages of gold deposits in

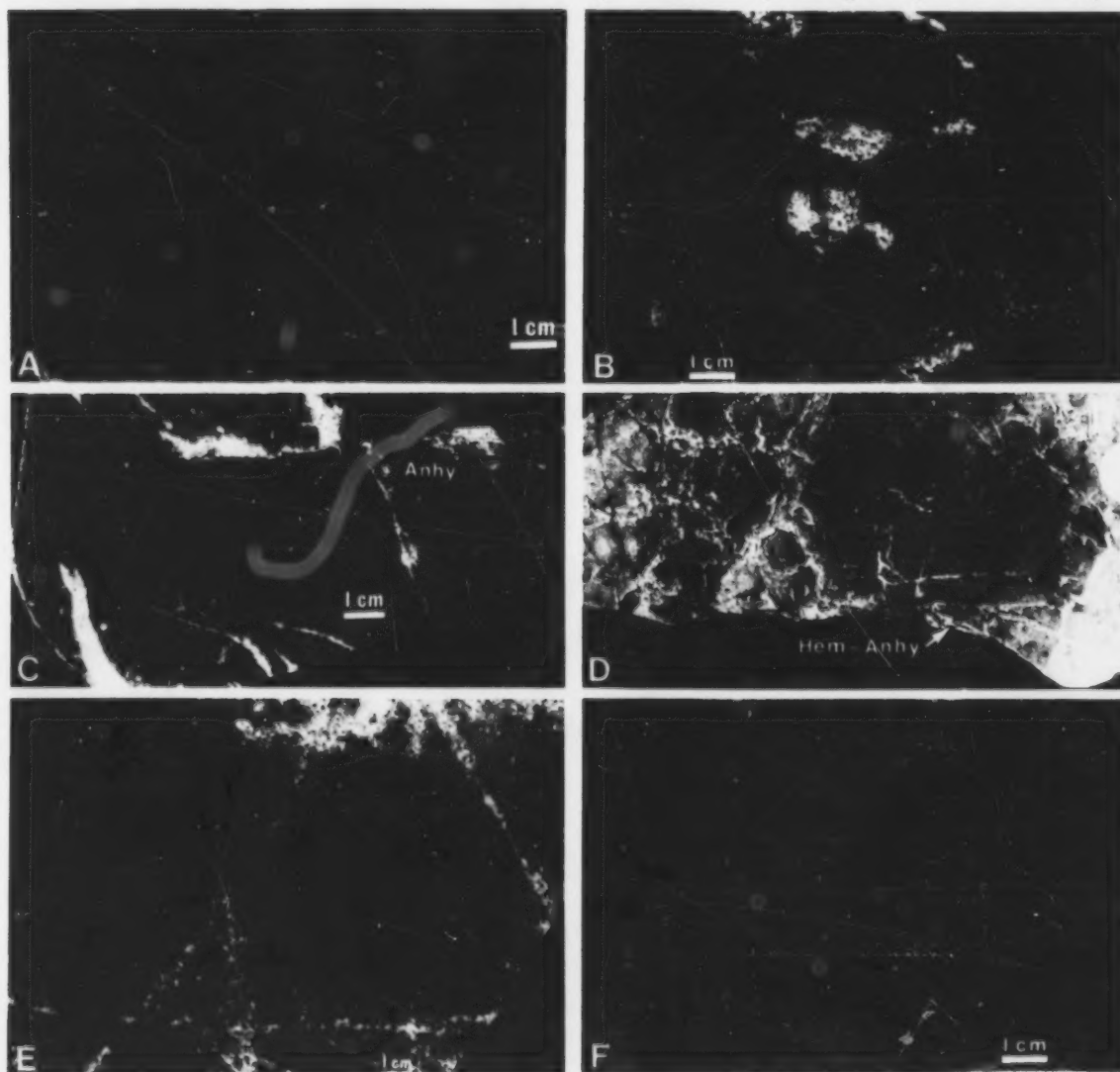
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different metamorphic domains also vary systematically, as illustrated in Figures 3.2 through 3.5.

Deposits hosted in greenschist grade rocks exhibit fracture-controlled, carbonate- and phyllosilicate-dominated alteration, associated with areas of most intense quartz veining. Examples of this style of alteration include the Hollinger and McIntyre (Keys 1940, Jones 1948, Mason and Melnik 1986), and Dome Mines (Holmes 1968, Fryer *et al.* 1979, Roberts 1987) in Timmins, the Kirkland Lake deposits (Thomson *et al.* 1948, Kerrich and Watson 1984), the Dickenson (A.W. White) Mine in Red Lake (Andrews and Hugon 1985), and the Golden Mile in Kalgoorlie (Boulter *et al.* 1987).

Conversely, alteration assemblages in deposits within amphibolite grade rocks are less hydrous, carbonate-poor and sulphide-rich. Alteration in these deposits is characteristically associated with a combination of vein and disseminated, replacement-type mineralization which often is localized at lithologic contacts. Some of the better examples include the amphibolite grade mineralization at the Campbell Red Lake and Dickenson (A.W. White) Mines in Red Lake (Andrews and Hugon 1985), the Musselwhite deposit at Opapimiskan Lake (Hall and Rigg 1986), and the Kolar Gold Fields of India (Hamilton and Hodgson 1986).

Supracrustal sequences around the Hemlo (Muir 1983, 1986) and Big Bell (Chown *et al.* 1984)



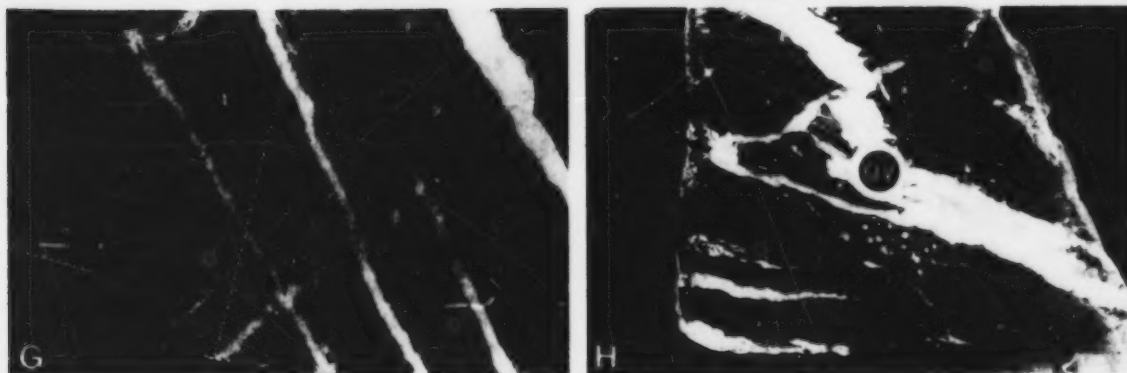


Plate 3.4. Texture and mineralogy of alteration in granitoid intrusives and porphyries. One centimeter scale bar provided.

- A. Sheared and sericitized intrusive porphyry. Croll Lake Stock, Geraldton.
- B. Highly schistose, siliceous and sericitic quartz-saturated intrusive porphyry. Pearl Lake Porphyry, McIntyre Mine, Timmins.
- C. Albitized (white crystals), mildly sericitized and carbonatized syenite intrusive rock with anhydrite (Anhy) fracture infillings. Feldspar is impregnated with fine hematite. From Ross Mine.
- D. Highly carbonatized and sericitized syenite porphyry adjacent to mineralization and contact with basalts. Hematite and anhydrite occur together as fracture infillings; epidote is also present in some fractures. From Ross Mine.
- E. Fracture control of alteration halos in medium grained siliceous quartz porphyry. Detour Lake Mine.
- F. Fracture control of alteration in a very fine grained and silicified intrusive porphyry. Detour Lake Mine.
- G. Trondhjemite cut by two sets of quartz veins, the latter of which exhibits hematite - sulphide alteration halos. Similar alteration occurs in fractures subparallel to the later quartz veins. Canadian Arrow Mine, Timmins.
- H. Feldspar porphyry cut by quartz - ankerite vein (QV) with halo of silicification and carbonatization extending from vein. Pyrite and green mica, after mafic minerals, occur in the zone of bleaching. From Hard Rock Mine, Geraldton.

deposits are metamorphosed to a higher grade than those associated with any other known deposits. These deposits exhibit a generally disseminated, foliation-parallel mineralization (Chown *et al.* 1984, Burk *et al.* 1986, Kuhns *et al.* 1986, Phillips and De Nooy 1988).

The general lack of carbonatization and the high degree of recrystallization in deposits within amphibolite grade terranes produces an appearance greatly different from that of alteration in greenschist grade rocks; hence, the identification of gold-related alteration in these terranes can be difficult. Metamorphic decarbonation of altered rocks around amphibolite grade gold deposits, as has been proposed by Groves and Phillips (1987), seems unlikely since the observed mineral assemblages are not necessarily consistent with decarbonation. Calc-silicate minerals, which are the common products of decarbonation of impure limestones (eg. diopside, enstatite, tremolite, wollastonite, grossular and epidote) should be present if decarbonation occurred, but are generally absent from gold deposits hosted

in amphibolite grade rocks. Exceptions are the occurrence of a hedenbergite - grossular - epidote assemblage surrounding veins at the Lupin Mine (Lhotka and Nesbitt 1987), and a diopsidic alteration at the Kolar Gold Fields (Hamilton and Hodgson 1986). Scapolite, a mineral commonly found in high grade calcareous rocks, has been identified at Hemlo (T.L. Muir, Geologist, Ontario Geological Survey, personal communication, 1988).

3.3.3.2. Timing of Metamorphism

Evidence from many deposits suggests a close temporal relationship between peak metamorphism, deformation, alteration, and the introduction of gold (see Chapters 1, 2 and 5, Part I). The best evidence that mineralization did not occur significantly before peak metamorphism comes from the overprinting by "retrograde" mineral assemblages in zones of most intense alteration, deformation and gold deposition. These assemblages, which are consistent with greenschist facies pressure - temperature conditions, may indeed have resulted from true ret-

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regression (ie. they replaced pre-existing peak metamorphic minerals as metamorphism waned (Winkler 1974)). An alternative explanation is that they formed at pressure - temperature conditions above their accepted field of stability. This hypothesis is discussed in the following section.

Examples of deposits where these apparently lower grade "retrograde" assemblages coincide with mineralization in higher metamorphic grade rocks include the mines at Red Lake (Andrews *et al.* 1986) and Detour Lake (Marmont 1986) and the Duport deposit (Smith 1986) in Ontario, and the Victory Mine at Kambalda (Clark *et al.* 1986). Significantly, in these examples, alteration is not necessarily limited to the zone of lower metamorphic grade mineral assemblages. At the Campbell Red Lake and Dickenson (A.W. White) Mines in Red Lake, which occur in amphibolite grade rocks, a predominantly muscovite + chlorite + biotite (ie. greenschist) assemblage occurs in altered basaltic wall rocks adjacent to gold mineralization. Peripheral to this assemblage is an apparent amphibolite grade assemblage of cordierite + staurolite + chloritoid + andalusite (Mathieson and Hodgson 1984, Andrews and Hugon 1985). The latter assemblage, while consistent with the regional amphibolite grade, is not the expected assemblage in basalts. Therefore, these rocks must have been hydrothermally altered. This juxtaposition of alteration and metamorphic assemblages points to the close temporal relationship of peak metamorphism and gold-related alteration. This relationship is discussed elsewhere in this report.

3.3.3.3. Mineral Assemblages, Fluid Composition and Fluid Flux

The mineral reactions that occur during progressive metamorphism are responses to heat flux, pressure - volume transitions, and interactions of rocks and fluids which are not in equilibrium (Ferry 1983a, b, c, 1986, Wood and Graham 1986). It has been demonstrated that, under some conditions, these reactions can to a large extent control both the temperature of metamorphism and the composition of metamorphic fluids (Ferry 1983c). However, in zones of high fluid flux, the temperature and composition of the fluid can control which reactions occur and, therefore, the mineral assemblages in these zones may not necessarily reflect the ambient pressures and temperatures of regional metamorphism.

One example of the control of fluid composition on mineral assemblages is the Victory Mine, Kambalda. Clark *et al.* (1986) suggested that high XCO_2 in the fluid phase controlled carbonation reactions in the Victory Mine area, resulting in assemblages characteristic of lower metamorphic grade than the regional assemblage. In shear zones along with the fluids were channeled, the lack of mineral assemblages containing both reactants and products of the carbonation reactions indicates that the fluid was not significantly buffered by the mineral assemblages of the surrounding rock.

The Ti-bearing minerals - ilmenite, sphene and rutile - are particularly sensitive to compositional gradients (Hunt and Kerrick 1977). A zonal distribution of sphene and rutile is observed at the Victory Mine, where rutile occurs in association with zones of most intense alteration and gold mineralization (Clark *et al.* 1986). Similar relationships in Ti-bearing minerals have been observed in Canadian deposits (eg. Sigma Mine, Robert *et al.* 1983; Ross Mine, Detour Lake Mine and Duport deposit, Troop *et al.* 1988). Other minerals in which zonation adjacent to mineralization has been documented are Ca - Fe - Mg - Al silicates (Brown 1975, Laird 1980, Moody *et al.* 1983) and carbonates (Ferry 1983a, b, c).

In contrast to assemblages which are controlled by fluid composition, studies of massive sulphide deposits have shown that sulphide and oxide mineral assemblages can buffer $f\text{O}_2$ and $f\text{S}_2$ during metamorphism (Nesbitt and Kelly 1980, Nesbitt 1982, 1986, Bryndzia and Scott 1987a). These suggested controls of mineral compositions and stabilities are corroborated by experimental results of sulphidation and oxidation in amphiboles (Popp *et al.* 1977a, b), biotites (Tso *et al.* 1979), and chlorites (Walshe 1986, Bryndzia and Scott 1987b). These conditions, however, may not relate to gold-associated alteration, which is characterized by higher fluid:rock ratios and smaller proportions of sulphide.

Studies of the mineral assemblages associated with mineralization are required to understand the nature of alteration reactions and to assess the relative control of fluid composition on mineral assemblages, or *vice versa*. Once these conditions are constrained, the relationship of mineral reaction surfaces to structure and mineralization may provide information on chemical gradients associated with alteration fluids.

3.4 SUMMARY

Alteration associated with Archean lode gold deposits displays consistent characteristics, which vary with the setting of the deposit. These characteristics include:

(1) *a regional component*: Alteration is associated with linear zones of intense deformation or crustal discontinuities which contain gold mineralization, and which may transgress greenstone - granitoid boundaries. Alteration occurs sporadically within these deformation zones.

(2) *a late cross-cutting relationship*: Alteration affects all Archean lithologies and all but the latest structural features.

(3) *a variation in style*: Alteration style varies with the style and degree of deformation and with metamorphic grade.

(4) *a characteristic suite of minerals, including a regional and local association with carbonates*: Carbonates occur both in wall rock alteration and in veins. Other characteristic minerals include potassic phyllosilicates, alkali feldspars, chlorite, Fe sulphides and quartz.

(5) *a progressive dilution of original bulk chemistry*: Alteration is accomplished by stripping of CaO and Na₂O, relative enrichment of Al₂O₃, and addition of K₂O, CO₂, H₂O, S, Au, and other incompatible elements.

The heterogeneity observed in alteration from deposit to deposit can be systematically explained in terms of three fundamental variables:

(1) *the bulk composition of the rock through which altering fluids pass*: Host rock lithologies limit the possible mineral:fluid reactions, and hence the resulting mineral assemblages. The effects of this control are diminished where very large volumes of fluid react with the rock.

(2) *the primary or deformation-enhanced permeability of the host rocks*: Permeability controls, in part, the extent to which a hydrothermal fluid may react with a heterogeneous assemblage of lithologies.

(3) *the P - T - fluid conditions of metamorphism*: The ambient conditions of metamorphism influence the minerals that form during the alteration.

Several questions remain to be addressed in refining the systematics of gold-related alteration. Chemical relationships of alteration are complicated by the heterogeneous geochemical and mechanical responses of host rocks during alteration. Further research into methods of normalization of whole rock compositions is considered pivotal in solving this problem.

Gold-related alteration assemblages have not been utilized effectively to provide information about the physical and chemical conditions of ore formation. To date, most data about these conditions have come from studies of stable isotopes and fluid inclusions (see Chapter 6, Part I of this report). One method of addressing this problem is to document the textural relations and compositional variations in alteration phases developed about gold-bearing structures. Such an approach has been taken at the Hunt Mine, Kambalda (Neall and Phillips 1987), where a thermodynamic model has been developed from compositional zoning in equilibrium silicate - sulphide - oxide assemblages in compositionally uniform basalts to describe fO_2 , fS_2 , and fluid composition during gold deposition.

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4. ASSOCIATION WITH FELSIC INTRUSIONS

4.1. INTRODUCTION

A close spatial association with granitic intrusive rocks is often cited in descriptions of Archean lode gold deposits, and has formed a part of many exploration strategies in the search for more deposits. This spatial association was discussed in Chapter 1 of this volume, where it was noted that post-volcanic, intra-belt plutons constitute more than 25% of host rocks to gold mineralization in the Abitibi greenstone belt, although they form less than 4% of the surface exposure of the belt. This chapter presents a more comprehensive description of the petrology and ages of these intrusions.

The anomalous spatial association between these syn- to late tectonic, felsic to intermediate plutons and gold deposits is illustrated on Figure 4.1. This association is in marked contrast to that displayed by the syn-volcanic, dioritic to trondhjemitic plutons which correspond in both age and composition with volcanic rocks in the greenstone belts (Paradis *et al.* 1988): they are not anomalously abundant in gold-mineralized areas. As well as displaying a spatial association with gold deposits, the syn- to late-tectonic plutons seem to occur more frequently in the deformation zones that host gold mineralization (Chapter 5). However, as discussed below, compositional variability and equivocal intrusive relationships often make this suite difficult to define without supplementary geochemical data.

4.2. PETROLOGICAL CHARACTERISTICS

Establishing the petrogenesis and relative timing of these intrusions is not always easy. Petrographic data, necessary for a comprehensive petrological characterization, are sparse. The limited data available indicate a wide range of silica-saturated compositions, from trondhjemitic to granodiorite to quartz monzonite, and occurrences of both equigranular and porphyritic textures. Porphyritic varieties contain quartz or quartz and feldspar phenocrysts. Many of the intrusions have been sufficiently altered (see Chapter 3) that identification of their original chemistry is not possible. However, trace element and REE geochemical signatures of some are sufficiently diagnostic that they can be classified as granodioritic to quartz monzonitic (eg. porphyries in the Timmins camp, Burrows and Spooner 1986).

In addition to the quartz-bearing plutons, a suite of silica-undersaturated intrusions hosts gold mineralization in the Timmins and Matheson areas of the Abitibi Subprovince. The undersaturated plutons, like the saturated, are sometimes intruded into deformation zones, but are smaller and less deformed than the latter. For example, a set of albitite (pebble) dikes, which host gold mineralization, cuts the quartz porphyries in the McIntyre and Hollinger Mines at Timmins. Burrows and Spooner (1986) have proposed that the shear-induced, prolate fabric of the porphyries is much more pronounced than that of the albitite dikes. Similarly, monzonitic intrusions in the Matheson area are only deformed along discrete high strain zones, and have for the most part retained pristine intrusive textures. Thus, these undersaturated plutons were emplaced during the later stages of the deformation.

The undersaturated intrusions have been intensely altered, as are all rocks in gold mineralized areas. However, rock geochemistry and remnant intrusive textures of several plutons suggest compositions ranging from syenite (Young - Davidson Mine (Sinclair 1982), Macassa Mine (Watson and Kerrich 1983)) to monzonite (Murphy - Garrison and Canadian Arrow Mines (Cherry 1983)) and lamprophyres (McIntyre and Hollinger Mines (Burrows and Spooner 1986)).

Preliminary observations suggest that some of the gold-related intrusions in the Abitibi belt may have evolved under relatively oxidizing conditions, as manifest by the presence of oxidized biotite and hornblende and the occurrence of sulphate minerals and hematite as minor phases away from zones of obvious alteration (eg. Cameron and Carrigan 1987). Precipitated sulphide minerals in gold ores within or spatially associated with these intrusions also have distinctive sulphur isotopic values (Schwarcz and Rees 1985, Cameron and Hattori 1987). While the premise that some gold-related intrusions may have evolved under relatively oxidizing conditions is intriguing, the data are too few to assess how widespread the oxidized magmas were. No equivalent data are available for synvolcanic intrusion, or for intrusions in areas away from gold mineralization.

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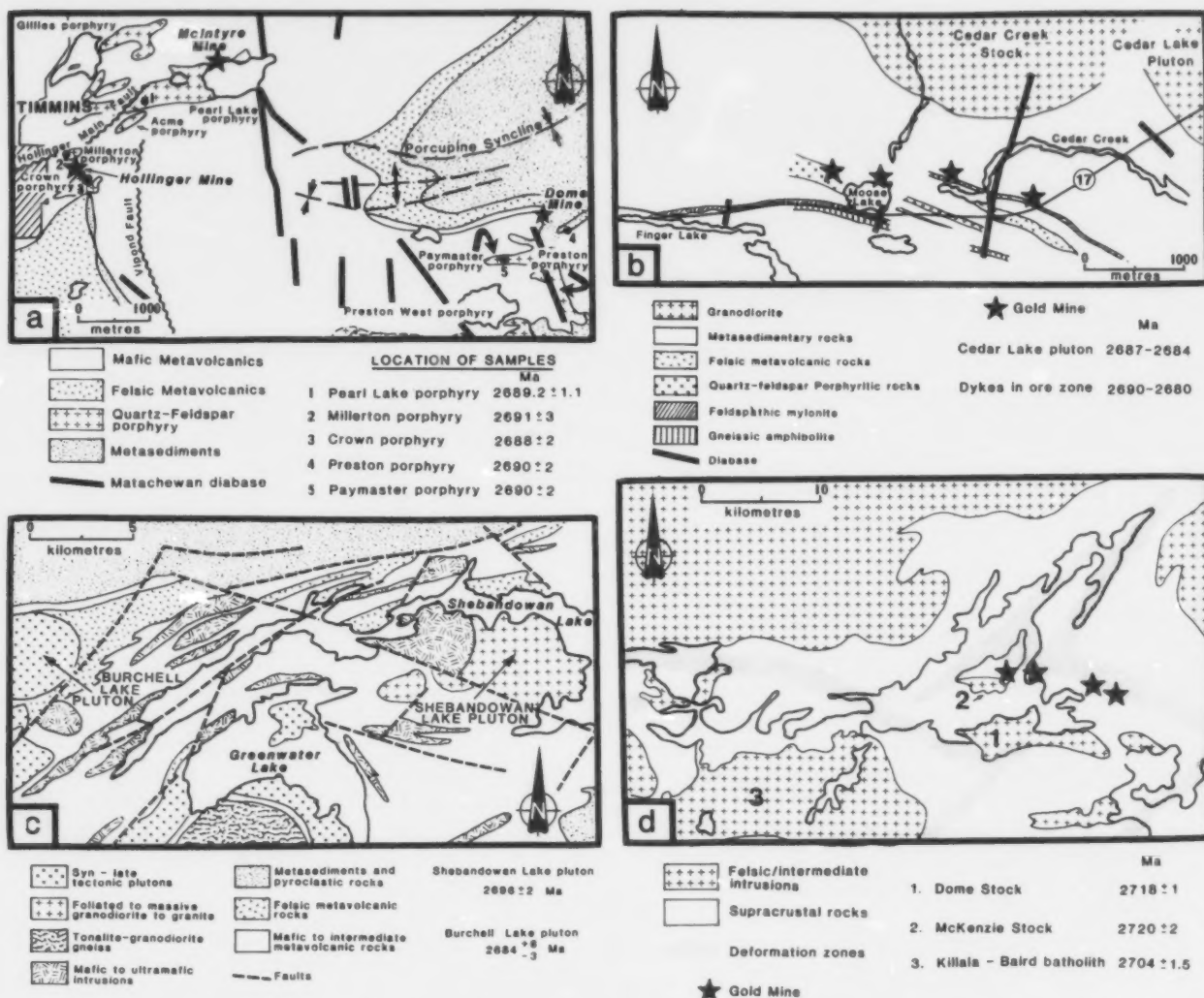


Figure 4.1. U/Pb zircon geochronology of post-volcanic, syn- to late tectonic, felsic to intermediate intrusions at: (a) Timmins, (b) Hemlo, (c) Shebandowan, and (d) Red Lake.

4.3. AGE OF INTRUSION

The intrusive contact relationships of many of the gold-associated plutons have been obliterated by shearing (eg. the McIntyre Mine in Timmins, Mason and Melnik 1986): in fact, many have recorded much of the shearing, and are highly schistose. This deformation makes establishing the time of their intrusion relative to volcanism, and folding and tilting of the surrounding strata difficult. Precise radiometric age determinations offer a method of determining the absolute ages of the intrusions, and hence interpreting their tectono-magmatic setting.

U-Pb zircon geochronology of gold-associated quartz-bearing plutons from gold camps in the Superior Province indicates a very tight cluster of ages. Marmont and Corfu (1988) obtained ages from 2691 to 2688 Ma. from plutons in the Timmins area (Figure 4.1a). Corfu and Muir (1988) bracketed the age of felsic intrusion, both within and adjacent to the ore zones at Hemlo (Figure 4.1b), between 2690 and 2670 Ma. Corfu and Stott (1987) reported ages of 2696 to 2686 Ma. for intrusions in the Shebandowan area in northwestern Ontario (Figure 4.1c). All of these plutons, which are located in the southeastern terranes of the Superior Province (see Chapter 1), were intruded in the period 2690 to 2670 Ma. This post-dates the

cessation of calc-alkaline volcanism in these greenstone belts by 10 to 30 Ma. In the Red Lake camp (Figure 4.1d), within the older northwestern terrane of the Superior Province, similar felsic intrusions returned ages of 2720 to 2700 Ma. (Corfu and Andrews 1987), which post-dates volcanism by a minimum of 20 to 40 Ma.

The quartz-bearing intrusions may be the youngest host rock for mineralization in a gold deposit. Colvine *et al.* (1984) and Mason and Melnik (1986) suggested that because of their strong spatial and temporal association with gold, the intrusions may have a genetic link with gold. However, recent geochronological work by Marmont and Corfu (1988) has locally refined the syn- to late-tectonic timing of the plutonism, and indicated that introduction of gold was distinctly later than the crystallization of at least the quartz-saturated suite of intrusions.

To date, very few of the undersaturated intrusions have been dated by the U-Pb zircon method. An albitite dike from the McIntyre Mine at Timmins returned an age of $2673 \pm 6/-2$ Ma., and a monzonite from the Murphy - Garrison Mine near Matheson returned an age of 2678 ± 2 Ma. (Marmont and Corfu 1988). These ages correspond to a sphene age of 2674 ± 2 Ma. obtained from a lamprophyre dike in the Adams (iron) Mine near Kirkland Lake (Wyman and Kerrich 1987). Thus, in the Timmins and Kirkland Lake areas, the undersaturated plutons, which are the youngest Archean intrusions, post-date volcanism by at least 25 Ma., and post-date intrusion of the quartz-bearing plutons by 15 Ma.

In the Hemlo area, a number of intrusions that occur in the ore zones have returned ages of 2690 to 2670 Ma. However, the intense deformation and alteration of these rocks prevents distinguishing between quartz-phyric and feldspar-phyric intrusive suites.

4.4. SUMMARY

Gold deposits in the Superior Province have a close spatial association with syn- to late-tectonic felsic intrusions. Although these intrusions are typically altered and deformed, it has been possible to identify two suites: an earlier, syn-tectonic suite of quartz-bearing plutons, and a later, late-tectonic suite of silica-undersaturated plutons. These intrusive suites are remarkably similar in absolute ages, as determined by U-Pb zircon geochronology. These precise age determinations have also established a chronological distinction between the two intrusive suites in the Timmins - Kirkland Lake area of the Abitibi belt, where the older, quartz-bearing plutons have ages of 2691 to 2688 Ma., and the undersaturated plutons have ages of 2678 to 2673 Ma.

Colvine *et al.* (1984) suggested that the older, quartz-bearing intrusions might be a source of gold-mineralizing fluids. While this is still possible, the role of the silica-undersaturated intrusions, as a source of fluids or as a manifestation of the gold-mineralizing process, must also be considered. Understanding the true role of these intrusive suites in the gold deposits still awaits the acquisition and interpretation of adequate petrological and petrochemical databases, which will permit more accurate classification and petrogenetic interpretations. However, despite these difficulties, the close spatial association of mineralization with a temporally distinct suite of intrusions remains an important empirical characteristic of Archean lode gold deposits. The genetic implications of this association are discussed in Part II of this volume.

PART I. DEPOSITIONAL MODEL

5. TIMING OF MINERALIZATION

5.1. INTRODUCTION

Determining when mineralization occurred, relative to the protracted evolution of the greenstone belts which host the lode gold deposits, is critical to understanding their genesis. Empirical evidence pertinent to this relative age of mineralization has been included in the broader discussions of the preceding chapters. This evidence is briefly recapitulated in the following section, as a preface to a discussion of the results of geochronological studies of the absolute age of the gold mineralization.

5.2. RELATIVE AGE OF MINERALIZATION

Relative age brackets for the maximum and minimum ages of gold mineralization across the Superior Province can be defined from geological relationships observed in the field. These observations include relationships between gold mineralization and:

(1) *host rocks* (see Chapter 1): Gold mineralization can be hosted by any lithology in a greenstone belt, and occasionally occurs in gneissic terranes outside greenstone belts. The oldest unequivocally post-ore units are Early Proterozoic mafic dikes. However, since gold occurs in all lithologies (and all ages of lithologies) in the greenstone belts, its lithological associations do not discriminate between two extreme cases: that the gold could have been introduced at any time over a protracted period (Hutchinson 1987), or that it was introduced after the formation of its youngest host (Colvine *et al.* 1984). This discrimination is only possible after consideration of evidence provided by studies of structures and alteration associated with the gold deposits.

(2) *deformation* (see Chapter 2): The deformation zones that host gold mineralization are of regional and Shield-wide extent, and display consistent orientations and sense of shear. The ubiquitous and systematic association of the mineralization with these discrete tectonic structures is compelling evidence that gold concentration was related to the major, latest Archean tectonic event that formed them.

Strain analyses of gold deposits have demonstrated that the present sites of mineralization represent zones of dilatancy pro-

duced by shear deformation. Structural studies also indicate that the shear fabric is superimposed upon fabrics produced by folding or tilting. The youngest greenstone lithologies that host gold deposits (the Timiskaming-type sediments, volcanics and associated intrusions) record only the second (shear) fabric. Thus, mineralization was synchronous with, or post-dated, shearing.

These deformation-related characteristics do not by themselves eliminate the possibility that reconcentration of earlier deposited gold occurred during this tectonism to form the present structurally-sited deposits. However, evidence derived from studies of the alteration associated with the gold deposits does provide additional constraints on timing.

(3) *alteration* (see Chapter 3): Gold deposits are typically surrounded by a large volume of altered rock. This alteration could only be effected by the passage of a large volume of fluid through the rock: this volume of fluid could not be locally derived. Zoned alteration patterns around both individual ore shoots and entire deposits indicate that the present site of mineralization is essentially its original depositional site, and argue strongly against remobilization of earlier gold deposits during deformation.

(4) *metamorphism* (see Chapter 3): Alteration zones around metamorphosed mineral deposits contain mineral assemblages diagnostic of the metamorphism. Good examples of this metamorphic overprinting are found in the anthophyllite + cordierite assemblage around many Archean volcanogenic massive sulphide deposits (Franklin *et al.* 1981, see also Chapter 3 of this report). The alteration zones around Archean lode gold deposits, however, show no evidence of any metamorphic overprint. Instead, the alteration mineral assemblages vary with the regional metamorphic grade. Furthermore, the alteration assemblages of deposits hosted in amphibolite facies rocks contain minerals which are stable at lower than estimated peak metamorphic temperatures: this is manifest as retrogression of metamorphic minerals by alteration minerals. Both lines of evidence, therefore, indicate that the hydrothermal activity of the gold mineralizing process occurred during and

PART I. DEPOSITIONAL MODEL

somewhat after the attainment of peak metamorphic conditions.

(5) *felsic intrusions* (see Chapters 1 and 4): Post-volcanic, syn- to late-tectonic felsic to intermediate plutons are anomalously abundant hosts to gold mineralization, relative to their abundance in greenstone belts. As the youngest host lithology to gold, they define the maximum relative age of mineralization.

The empirical evidence presented above indicates that the gold was introduced in a single event that followed the intrusion of syn- to late-tectonic felsic plutons into folded and sheared volcano-sedimentary sequences in the greenstone belts. A more accurate estimate of when the mineralization occurred in the Archean can be made from absolute ages obtained from precise geochronology, accepting the premise that mineralization is younger than any lithology that hosts it.

5.3. ABSOLUTE AGE OF MINERALIZATION

The geochronology of volcanism and plutonism in greenstone terranes across the Superior Province in Ontario was discussed in Chapter 1, Part I of this volume, in order to provide an understanding of the regional geological setting of the gold deposits. Although specific dates are not repeated here, those geochronological data show that:

(1) The lithological units hosting mineralization can encompass the entire range of ages obtained from a greenstone belt (Figure 1.5). For example, in the Red Lake camp, gold is hosted by rocks that formed over a period of 272 Ma. (Corfu and Andrews 1987). Similarly, in the Abitibi belt, gold is hosted by rocks that formed over a period of 80 Ma. (Marmont and Corfu 1988).

(2) The maximum age of mineralization obtained for the northwestern greenstone terranes is 2718 Ma. (Figure 5.1), from the Red Lake camp (Corfu and Andrews 1987). This post-dates the youngest volcanism in the Red Lake camp by approximately 15 Ma.

(3) In the southeastern greenstone terranes,

the maximum age of mineralization is 2690 to 2673 Ma. (Figure 5.1), which post-dates the youngest volcanism by approximately 10 to 25 Ma. In the Timmins area, geochronology has shown that the late-tectonic, silica-undersaturated felsic plutons post-date the quartz porphyries by 13 to 15 Ma. Gold, hosted in both types of pluton, must be synchronous with or younger than the undersaturated suite.

(4) In the Wabigoon Subprovince, the range of ages obtained for the youngest, post-volcanic, felsic intrusions is 2709 to 2695 Ma. These dates are older than, or equivalent to, the ages of the quartz porphyries in the Timmins camp (2691 to 2688 Ma.). More importantly, a suite equivalent to the younger (2678 to 2673 Ma.), undersaturated intrusions of the Timmins camp may not have been recognized.

The difference in the maximum age of the introduction of gold in the northwestern and southeastern greenstone terranes suggests that the same mechanisms of mineralization were diachronously active in the final stages of cratonization of older and younger Archean crusts across the Superior Province. This suggestion, and the possible genetic link between gold mineralization and the undersaturated intrusions, are discussed in Part II of this report.

The minimum age of mineralization has not, as yet, been well defined. To date, no lithology other than Lower Proterozoic diabase dikes has been unequivocally shown to cut mineralization. The gap between the crystallization age of the dikes and the age of the youngest host lithology is too great to provide a meaningful bracket for the duration of mineralization.

A minimum age of mineralization has been suggested for the Red Lake camp by Andrews *et al.* (1986). Pre- to syn-deformation intrusions at Red Lake date at about 2720 Ma, and late to post-deformation dikes date at 2700 Ma. Andrews *et al.* (1986) noted that, since gold mineralization at Red Lake is spatially and temporally related to the deformation bracketed by these two ages, it too is bracketed, and must have occurred between 2720 and 2700 Ma.

In the Hemlo area, Corfu and Muir (1988) report U-Pb ages of 2650 to 2610 Ma. for rutile

and monazite, which are presumed to be part of the hydrothermal assemblage and possibly synchronous with mineralization.

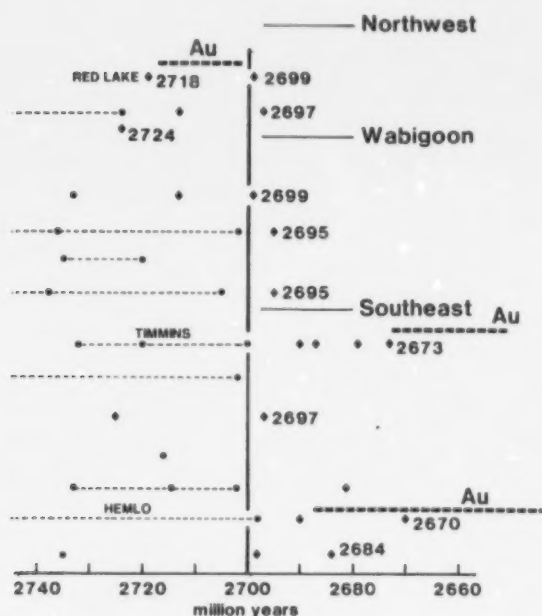


Figure 5.1. U/Pb geochronological data for latest volcanics (circles) and syn- to late tectonic plutonic (diamonds) rocks in the northwestern, Wabigoon and southeastern greenstone terranes of the Superior Province. See Section 4, Chapter 1, Part I for further explanation.

Attempts have been made to determine the age of mineralization in the Abitibi belt by dating hydrothermal vein minerals that occupy the same structures as gold and are considered synchronous with it. A few radiometric determinations have been made on green micas from the veins and wall rocks of the Dome and Hollinger Mines, using the $^{40}\text{Ar}/^{39}\text{Ar}$ method (Masliwec *et al.* 1986). The ages obtained range from 2633 to 2617 Ma. When combined with the U-Pb ages discussed earlier, this indicates that mineralization at the Hollinger Mine occurred between 2673 and 2633 Ma. However, it should be emphasized that the $^{40}\text{Ar}/^{39}\text{Ar}$ method is best used to determine cooling histories, rather than crystallization ages, of minerals. Thus, establishing the age of formation of the hydrothermal mineralogy associated with gold mineralization is still under examination.

5.4. SUMMARY

Geochronological studies have shown that

the cessation of volcanism and the maximum possible age of gold mineralization in the greenstone terranes of the northwestern Superior Province are distinctly older (by 25 Ma.) than in the southern part of the Province. Regardless of these possible differences in maximum age of gold mineralization, and the difficulty of obtaining an accurate minimum age of mineralization, the geochronological data are consistent with the interpretations made from studies of structure, alteration and metamorphism associated with the Archean gold deposits. All of this evidence indicates that gold introduction occurred during the latest Archean, and at essentially the same time across extremely large areas of the Superior Province.

PART I. DEPOSITIONAL MODEL

6. EVIDENCE FOR FLUID TYPE

6.1. INTRODUCTION

The model developed in the preceding chapters involves the introduction of large volumes of fluid upward into the depositional environment along structurally induced, permeable channel ways. The systematic consideration of the effects of the interaction between that fluid and rocks in the depositional environment is based on the working hypothesis that fluids of grossly similar compositions were active in most, if not all, gold mineralizing systems. While systematic patterns of alteration are consistent with the premise that they were the products of a universally similar hydrothermal fluid, more direct evidence of the nature of the fluid is provided by application of fluid inclusion and stable isotopic techniques. Some results from typical studies are reported in this chapter, with emphasis on Canadian and Western Australian Archean gold deposits. Rigorous interpretations of these data as they pertain to source of the fluids and transportation of gold are deferred to Part II.

6.2. FLUID INCLUSION CHARACTERISTICS

An increasing number of studies of fluid inclusions in Archean lode gold deposits in the Superior Province are available. Such studies have been reported for the Doyon (Guha *et al.* 1982), Sigma (Robert and Kelly 1987) and O'Brien (Krupka *et al.* 1977) mines in Quebec; and in Ontario for the Timmins-Porcupine (Smith *et al.* 1987, Walsh *et al.* 1986, Wood *et al.* 1984, 1986) and Red Lake (LaKind and Brown 1984, Brown and LaKind 1986) camps; the Sioux Lookout - Sturgeon Lake (Brown 1985) and Geraldton (Macdonald 1984b) areas; the Wabigoon Subprovince (Brown 1986), and the Renabie Mine near Wawa (Studemeister and Kilias 1987). Some of the results from these studies are summarized in Table 6.1. As a general cautionary note, few studies demonstrate that gold occurs either in an inclusion, or along a fluid inclusion-filled fracture plane (eg. Robert and Kelly 1987). Therefore, it is not always possible to state conclusively which fluid inclusion population attended transport and precipitation of gold. Further, in many deposits, fluid paragenesis is not certain, because cross-cutting relationships between inclusion populations are not observed.

The fluid interpreted to have been trapped during the precipitation of gold was of low salinity (< 6 wt % NaCl equivalent), aqueous, but CO₂-bearing, with a moderate to high CO₂ density (0.7-

>1.0 g/cm³). Homogenization temperatures for the fluid inclusion data range between 400°C and 200°C, and cluster around 350°C. Because the homogenization temperatures provide only a minimum temperature, the actual trapping temperature may have been higher. Calculated and estimated trapping pressures, determined from isochores for mixed H₂O-CO₂ inclusions, lie in the range of 1.5 to 4.5 kb (Smith *et al.* 1984, Brown 1986, Brown and Lamb 1986, Robert and Kelly 1987), which corresponds to a depth of 4 to 12 km, assuming lithostatic load. These inclusions may also contain minor amounts (<5 volume %) of CH₄ (LaKind and Brown 1984, Smith *et al.* 1984, Brown 1985, Brown and Lamb 1986, Wood *et al.* 1986). Variable CO₂/H₂O ratios in coexisting fluid inclusions associated with gold mineralization are observed in some deposits (Guha *et al.* 1982, Macdonald 1984a, Smith *et al.* 1984, Brown and LaKind 1986, Wood *et al.* 1986, Robert and Kelly 1987). In the Sigma and Hollinger Mines, spatial relationships between two compositionally different coexisting fluid inclusion populations (eg. saline, aqueous fluid with minor CO₂ vs. low salinity, CO₂-rich fluid with minor H₂O and CH₄) suggest the two fluids were produced by unmixing of an initially homogeneous H₂O - CO₂ fluid of low to moderate salinity, containing approximately 5 to 30 mole percent CO₂ (Robert and Kelly 1987, Wood *et al.* 1986). The "reconstitution" of the auriferous fluid is based on the assumption that the wetting characteristics of the saline aqueous and the dilute CO₂-rich components are approximately equal, and that they had equal affinity to be trapped within growing quartz.

Fluid inclusion data on Archean vein-type gold deposits throughout the world (Roedder 1984, English 1981, Ho *et al.* 1985) are similar to those reported for the Superior Province. Thus, Archean auriferous fluids appear to be characterized by high fluid pressure, low salinity and presence of CO₂.

6.3. LIGHT STABLE ISOTOPE CHARACTERISTICS

Described in this section are selected, representative stable isotopic data for some Archean gold deposits. A more rigorous discussion of these data, with implications for fluid sources and generative process, is presented in Part II of this volume. A striking characteristic of these data is the similarity between Canadian and Australian Archean gold systems - a testament to the common source and process involved in their generation.

PART I. DEPOSITIONAL MODEL

DEPOSIT OR AREA (Reference)	FLUID INCLUSION DATA	INTERPRETATION
Red Lake, Ontario Brown and LaKind (1986)	Primary, low salinity H ₂ O-CO ₂ ; higher CO ₂ density; Variable CO ₂ /H ₂ O ratios; T _h = 110° to 360°C; GOLD RELATED. Secondary aqueous brine; T _h = >350°C. Secondary low salinity (<10 equi. wt % NaCl), CO ₂ - H ₂ O; low CO ₂ density.	Possible fluid unmixing
McIntyre-Hollinger Mine, Timmins, Ontario Smith et al. (1984)	Pseudo-secondary or primary H ₂ O-CO ₂ , T _h = 220° to 385°C, 3 to 24 mole wt % CO ₂ , CO ₂ liquid phase homogenization, CO ₂ density range 0.46 to 0.78 g/cm ³ ; GOLD RELATED. Secondary CO ₂ -bearing aqueous inclusions, T _h = 160° to 215°C. In graphite-bearing vein, pseudo-secondary to primary CH ₄ -CO ₂ -H ₂ O inclusions (>26 mole % CH ₄). Secondary CH ₄ -rich inclusions.	Possible fluid unmixing; fluids buffered by QFM; depositional pressure 0.35 to 0.5 bars
Hollinger Mine, Timmins, Ontario Wood et al. (1986)	Primary H ₂ O-CO ₂ , variable phase proportions; 1 to 4 equi. wt. % NaCl; 6 mole % CO ₂ ; T _h = 225° to 355°C, mean 277°C; minor CH ₄ ; CO ₂ density 0.65 g/cm ³ .	Phase separation
O'Brien Mine, Cadillac, Quebec Krupka et al. (1977)	Primary H ₂ O-CO ₂ with constant phase proportions. Coexisting secondary H ₂ O-CO ₂ ; T _h of both types = 210° to 380°C	Early fluid homogeneous Late heterogeneous fluid
Sigma Mine, Val d'Or, Quebec Robert and Kelly (1987)	Secondary H ₂ O-CO ₂ inclusions with <10 equi. wt. % NaCl, 15 to 30 mole % CO ₂ ; T _h = 285° to 395°C. Coexisting H ₂ O-rich inclusions, 25 to 34 equi. wt. % NaCl, T _h = 60° to 295°C and CO ₂ -rich inclusions in healed fractures.	Fluid unmixing intermittently
Yilgarn Block, Western Australia Ho et al. (1987)	Primary H ₂ O-CO ₂ with both constant and variable phase proportions (20 to 30 mole % CO ₂), <2 equi. wt. % NaCl, T _h = 200° to 390°C. Coexisting primary H ₂ O-rich and CO ₂ -rich in late vugs.	Early fluid, some unmixing Late fluid, heterogeneous

Table 6.1. Summary of fluid inclusion data for Archean gold deposits in the Abitibi Subprovince of Canada and the Yilgarn Block, Western Australia.

6.3.1. CARBON ISOTOPES

Hydrothermal carbonate minerals in veins and wall rocks of auriferous systems are a conspicuous component of the gold-associated gangue mineralogy, which precipitated from the CO₂-bearing, hydrothermal fluid. From the carbon isotopic compositions of the carbonates, it may be possible to constrain the source of hydrothermal carbon, and perhaps identify possible precipitation mechanisms related to changes in pH, redox conditions, and temperature, because of the dependency of carbon isotope fractionation on these parameters. The results of studies of Archean lode gold deposits by Donnelly *et al.* (1977), Fyon *et al.* (1980, 1982,

1983a), Golding and Wilson (1983), Golding *et al.* (1987), Spooner *et al.* (1985), McNeil and Kerrich (1986) and Kerrich *et al.* (1987) are shown in Figure 6.1. A general chronology of hydrothermal carbonate introduction, recognized in the greenstone belts of the Superior Province, is shown in Figure 6.2. This chronology has been derived from many studies, including work by Bain (1933), Davies *et al.* (1982), Fyon *et al.* (1983a), Andrews *et al.* (1986), Smith (1986), and Melling *et al.* (1986).

6.3.1.1. Abitibi Subprovince - Superior Province

Calcite was introduced early during diagenesis and alteration of the volcanic succession in the

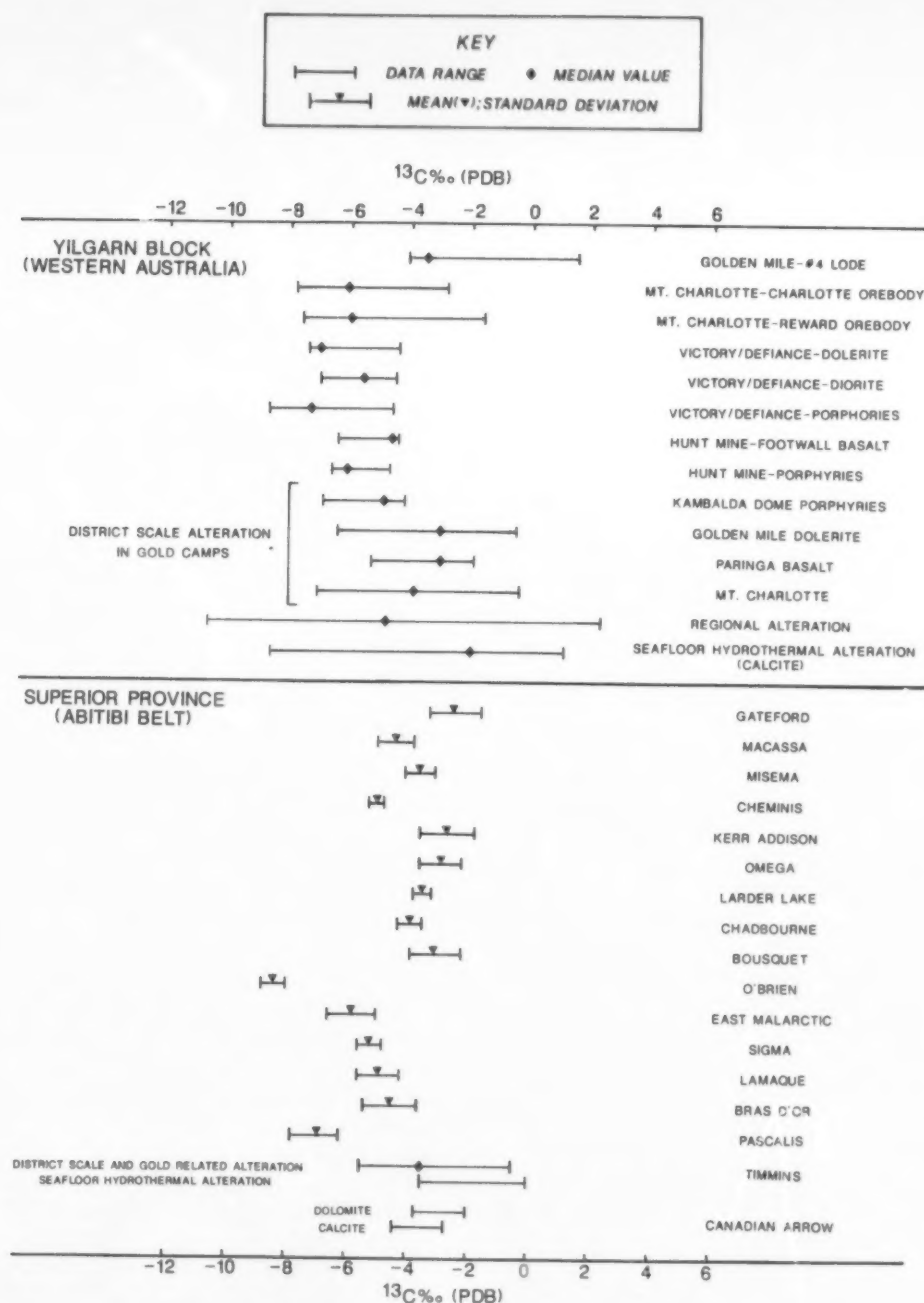


Figure 6.1. Partial compilation of $\delta^{13}\text{C}$ values of hydrothermal carbonate from the Abitibi Subprovince and the Yilgarn Block, Western Australia. Data for the Abitibi from Kerrich *et al.* (1987), Wood *et al.* (1986), Spooner *et al.* (1985), McNeil and Kerrich (1986), Fyon (1986), and Fyon *et al.* (1983a). Data for Western Australia taken from Lambert *et al.* (1984), Golding and Wilson (1983), and Golding *et al.* (1987).

Abitibi Subprovince, and occurs as a widely dispersed phase in greenstone lithologies, filling primary porosity (eg. Dimroth and Lichtblau 1979, Fyon *et al.* 1983a, b, Piroshco and Hodgson 1988).

The widespread distribution of the dispersed calcite, independent of regional structures, in volcanic rocks which have undergone low-temperature, sea water alteration (Beaty and Taylor 1979, 1982), suggests

PART I. DEPOSITIONAL MODEL

TIME	CARBONATE MINERAL	TEXTURE/HABIT	$\delta^{13}\text{C}$ (PDB)
OLDER  YOUNGER	CALCITE	DISSEMINATED, FILLING PRIMARY POROSITY. REGIONAL SEA FLOOR ALTERATION	0.0 TO -3.5‰ MODE: -1.7‰
	DOLOMITE-ANKERITE/ MAGNESITE → CALCITE (CHLORITE)	VEIN + REPLACEMENT. REGIONAL TO PROPERTY SCALE IN ZONES OF CARBONATIZATION.	-0.5 TO -5.0‰ MODE: -3.5‰
	DOLOMITE-ANKERITE/ MAGNESITE (SERICITE) (AU)	AURIFEROUS VEIN- RELATED CARBONATE.	MODE: -3.5‰(?)
	CALCITE	LATE, NON-AURIFEROUS QUARTZ VEINS.	NO DATA

Figure 6.2. Sequence of carbonate introduction into volcanic lithologies in the Timmins area, with corresponding carbonate mineralogy and carbon isotopic composition.

that the dispersed calcite may have been precipitated during the early sea floor diagenetic and alteration history of the flows. Dispersed calcite in the Timmins area has $\delta^{13}\text{C}$ values ranging from -3.5 to 0 per mil, with a mode at -1.7 per mil (Figure 6.1), consistent with a sea water source, given that the $\delta^{13}\text{C}$ values of calcite precipitated from Archean sea water show no secular change (Schidlowski *et al.* 1975, Veizer and Hoefs 1976).

Superimposed upon the early dispersed calcite are zones of intense carbonatization. These consist of dolomite - ankerite cores and chlorite - calcite margins, and may be centred on quartz - carbonate veins or shear zones (eg. Burrows 1912, 1924, Bain 1933, Fyon and Crocket 1982, Burrows and Spooner 1986, Wood *et al.* 1986, Piroshco and Hodgson 1988). This pattern of alteration developed in response to progressive fluid/rock interaction (eg. Clark *et al.* 1986). The zones of carbonatization are regionally distributed, and are related to late zones of deformation and zones of gold mineralization (Fyon and Crocket 1982, Davies *et al.* 1982). The $\delta^{13}\text{C}$ values of replacement dolomite - ankerite and magnesite (in ultramafic rocks) from within or adjacent to quartz veins range from -5 to 0 per mil, with a mode of -3.5 per mil (Figure 6.1). Within discrete zones of carbonatization, the $\delta^{13}\text{C}$ of carbonate in wall rocks varies by about 3 to 5 per mil across the zone, from about -4 per mil adjacent to a quartz vein in the dolomite-ankerite facies, to 0 per mil at the perimeter (chlorite +

calcite) of the zone. This variation is attributed to Rayleigh fractionation during progressive fluid/rock interaction (Fyon *et al.* 1983a, b, Fyon 1986).

Locally, a younger ankerite-dolomite + white mica replacement assemblage is superimposed upon the earlier dolomite-ankerite-calcite assemblages within a zone of carbonatization. This younger assemblage developed adjacent to auriferous quartz veins (eg. Bain 1933, Wood *et al.* 1986, Piroshco and Hodgson 1988). Both generations of ankerite-dolomite carbonate assemblages in the wall rock replacement zones are localized by the same late structures, which are also the structures into which gold was concentrated. The coexistence of gold and the two generations of ferroan carbonate, within the same late, regional structures, implies that the introduction of both CO_2 and gold was coeval with the late deformation event. The $\delta^{13}\text{C}$ values of this gold-related ferroan carbonate appear to be similar to that of the pre-gold dolomite-ankerite (-3.5 per mil; Spooner *et al.* 1985, Wood *et al.* 1986). Late, barren quartz-calcite veins were emplaced followed the main gold-ferroan carbonate introduction event (Bain 1933). No carbon isotopic data are available for this calcite.

An additional complication exists in the Timmins area. Reaction between the fluid and carbonaceous sediments may have induced isotopic fractionation between reduced and oxidized carbon-bearing fluid species, resulting in the precipitation of heavy, ^{13}C -enriched (up to 0 per mil) dolomite-

ankerite (Fyon *et al.* in preparation). Therefore, the $\delta^{13}\text{C}$ value of the total dissolved carbon in the hydrothermal fluid, before it encountered those domains containing carbonaceous sediment, was probably lower than the median value of -3.5 per mil. In fact, vein and replacement carbonates do show a general trend to heavier $\delta^{13}\text{C}$ values, from approximately -5 per mil in the outskirts of the Timmins district, where carbonaceous sediments are rare, to 0 per mil in the core of the Timmins camp, where carbonaceous sediments are present (Fyon *et al.* in preparation).

Thus, in the Timmins area, a 5 per mil variation in $\delta^{13}\text{C}$ of gold-related and regional carbonate is observed at two scales: camp scale (-5 to 0 per mil, and deposit scale (-3.5 to 0 per mil). The camp scale variation may reflect carbon isotopic fractionation between reduced and oxidized carbon species (Fyon *et al.* in preparation), while an additional element of variation on the deposit scale may reflect progressive depletion of CO_2 from the hydrothermal fluid (Rayleigh fractionation), by consumption of carbon during the precipitation of the dolomite-ankerite and chlorite + calcite assemblages (Fyon 1986). Thus, the lowest value (-5 per mil) may be a more accurate representation of the carbon isotopic value for carbonate which was precipitated from the hydrothermal fluids.

Kerrich *et al.* (1987) provide a large data base of carbon isotopic data which are representative of the Abitibi Subprovince. Average $\delta^{13}\text{C}$ values for specific areas range from -2.5 to -8.5 per mil, but are broadly consistent with the range of data reported from the Timmins area.

6.3.1.2. Yilgarn Block - Western Australia

In the Yilgarn Block, calcite which is interpreted to have been precipitated during an early, sea water alteration event, has a mean $\delta^{13}\text{C}$ value of -1.4 per mil (median -1.3 per mil; Golding *et al.* 1987), similar to that reported for the Timmins area. Carbonate from regional zones of carbonatization related to major fault zones, has a mean $\delta^{13}\text{C}$ value of -4.6 per mil (median -4.8 per mil; Golding *et al.* 1987). Carbonates from district scale zones of alteration have $\delta^{13}\text{C}$ values of -5.1 per mil (median -4.9 per mil), which is indistinguishable from the regional carbonate (Golding *et al.* 1987). Gold deposit-related alteration shows somewhat more variation in $\delta^{13}\text{C}$ values, from -7 to -3 per mil (Golding *et al.* 1987, Donnelly *et al.* 1977).

6.3.1.3. Summary

The similarity between the carbon isotope data from the Abitibi Subprovince and the Yilgarn Block is striking, and implies similar sources for the carbon. Under the indicated temperature (300-400°C), redox conditions (Smith *et al.* 1984, Kerrich *et al.* 1987), and pH conditions (neutral to slightly acid, where sericite stable), which approximate those of the auriferous fluid, the calculated $\delta^{13}\text{C}$ value for the total dissolved carbon in the hydrothermal fluid would be approximately that of the precipitated carbonate (Ohmoto and Rye 1979). Thus, the total dissolved carbon in the hydrothermal fluid would have had an approximate $\delta^{13}\text{C}$ of -7 to -0 per mil (Fyon *et al.* 1983a, Spooner *et al.* 1985, Kerrich *et al.* 1987, Cameron and Hattori 1987). Adoption of the median value value (-3.5 per mil) may provide a more realistic approximation of the $\delta^{13}\text{C}$ of the CO_2 component of the hydrothermal fluid, given the isotopic variation induced at the depositional sites, which produced ^{13}C -enriched carbonates.

6.3.2. SULPHUR ISOTOPES

Sulphide minerals (eg. pyrite, pyrrhotite, arsenopyrite, etc.) are ubiquitously associated with Archean gold systems, although their abundances vary within ore shoots in individual gold deposits. The sulphur isotopic compositions of these minerals provide a means to evaluate possible sources of sulphur and may reveal clues to depositional mechanisms, because of the sensitivity of sulphur isotopic fractionation to changes in redox conditions, temperature, and sulphur concentration in the fluid (Sakai 1968).

6.3.2.1. Canadian Archean

Sulphur isotopic data for Archean gold deposits in Canada (Figure 6.3) have been reported by Wanless *et al.* (1960), Lavigne (1983 1984), Schwarcz and Rees (1985), Spooner *et al.* (1985), and Cameron and Hattori (1987). Many of these studies have been reconnaissance in nature, and the significance of isotopic variation between different sulphide mineral morphologies or generations has not been exhaustively addressed.

Two broad groupings of $\delta^{34}\text{S}$ values are apparent: $\delta^{34}\text{S} = 0$ to +10 (Yellowknife, Red Lake, Belmoral, Owl Creek, Coniaurum, McIntyre-Hollinger, Dome, Hard Rock) and $\delta^{34}\text{S} < 0$ (Lakeshore, Macassa, Canadian Arrow, Kelore, Ross, Young Davidson, Consolidated Matachewan, Hemlo).

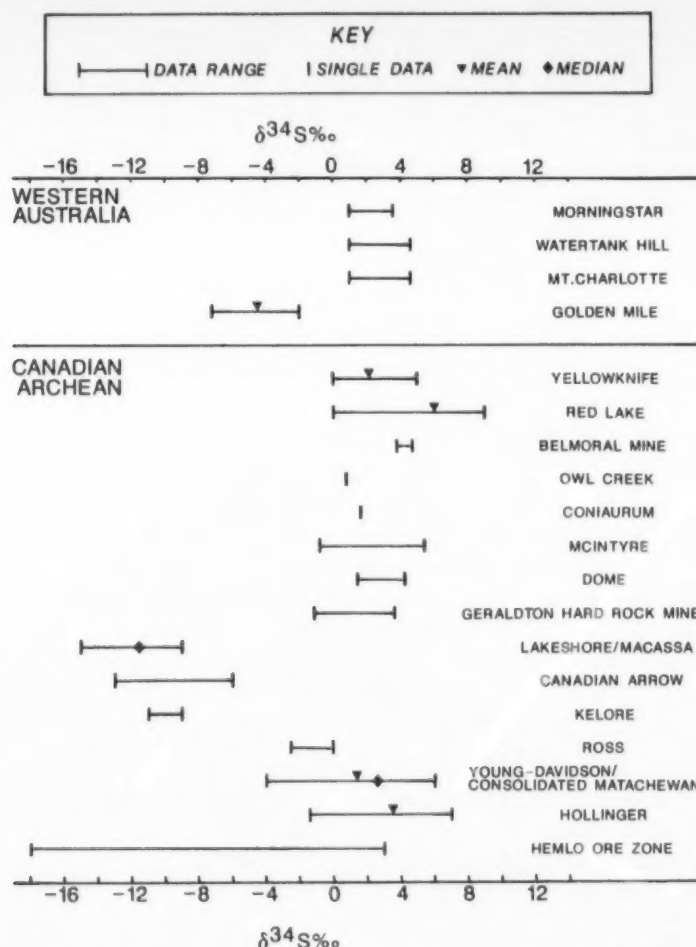


Figure 6.3. Compilation of $\delta^{34}\text{S}$ values of pyrite and pyrrhotite from Archean gold deposits. Data taken from: Western Australia - Lambert *et al.* (1984); Yellowknife - Wanless *et al.* (1960); Red Lake - Lavigne (1983, 1984); Geraldton (Hard Rock Mine) - Schwarcz and Rees (1985); Abitibi Belt (Belmoral, Owl Creek, Coniaurum, McIntyre, Dome, and Canadian Arrow Mines) - Schwarcz and Rees (1985); Abitibi Belt (Hollinger Mine) - Schwarcz and Rees (1985), Spooner *et al.* (1985); Abitibi Belt (Lakeshore, Macassa, Kelore, Ross, Young-Davidson, Consolidated Matachewan, and Hemlo deposits) - Cameron and Hattori (1987).

This division appears more transitional than abrupt (Figure 6.3), and 0 per mil is a convenient, albeit arbitrary, discriminant value. The low sulphur values (<0 per mil) may reflect relatively oxidizing hydrothermal fluids, as demonstrated by the presence of sulphate minerals and hematite as gangue minerals in the wall rock or within the auriferous systems (Cameron and Hattori 1987). Note, however, that sulphate minerals are present within the McIntyre mineralized system (Davies and Luhta 1978), although distinctly negative $\delta^{34}\text{S}$ values for pyrite are not typical of the Hollinger-McIntyre system. Many deposits display intra-deposit isotopic variation, suggestive of either mixing of more than

one sulphur source or major isotopic fractionation. Given the evidence for oxidizing conditions seen in some deposits, redox change within and between deposits, with attendant isotopic fractionation, is likely the dominant factor contributing to sulphur isotopic variation (Schwarcz and Rees 1985, Cameron and Hattori 1987).

6.3.2.2. Yilgarn Block - Western Australia

Lambert *et al.* (1984) present sulphur isotopic data for the Hunt, Morning Star, Mount Charlotte and Watertank Hill deposits. These data show a range from +0.7 to +8.0 per mil

(arithmetic mean of +3.0 per mil), which compares favourably with the data for that group of Canadian deposits whose $\delta^{34}\text{S}$ values are greater than 0 per mil. $\delta^{34}\text{S}$ values for pyrite from the Golden Mile lodes are notably lower, ranging from -9.0 to 0 per mil (arithmetic mean of -6 per mil; Lambert *et al.* 1984, Cameron and Hattori 1987). These negative values are similar to that Canadian population whose $\delta^{34}\text{S}$ values are <0 per mil (Figure 6.3). Sporadic anhydrite occurrences are reported in the lodes of the Golden Mile and in the Paringa Basalt (reported in Cameron and Hattori 1987), consistent with the suggestion that oxidized conditions existed during the precipitation of the pyrite.

6.3.2.3. Summary

The general similarity between types of sulphur isotope distribution representative of both the Superior Province and Western Australia emphasizes the probability of a uniform sulphur source, and a similarity in factors which contributed to the fractionation of sulphur isotopes and the precipitation of the sulphide minerals. While the large range in $\delta^{34}\text{S}$ data may suggest diverse sources of sulphur, sulphur isotopic partitioning between oxidized and reduced sulphur species in the hydrothermal fluid could also account for the spread in $\delta^{34}\text{S}$ values of precipitated pyrite (Schwarcz and Rees 1985, Cameron and Hattori 1987). A value of 0 per mil may best approximate the $\delta^{34}\text{S}$ of the total dissolved sulphur in the hydrothermal fluid.

6.3.3. OXYGEN ISOTOPIC DATA

Vein quartz is the most abundant constituent in most Archean gold deposits, and its oxygen isotopic characterization has been the focus of many studies. Oxygen isotope geothermometry is a secondary application of this technique; however, because mineral triplet data are not illustrated or frequently do not support isotopic equilibrium, the interpretation and geological significance of these calculated temperatures may not be straight-forward. In this regard, the broadest temperature constraints imposed by the oxygen isotope geothermometry are similar to those provided by other techniques (eg. fluid inclusion filling temperatures). Interpretations of these data and extrapolation to the conditions of gold precipitation must be done cautiously, because, in many cases, gold is presently situated in fractures which cross cut quartz crystals, and therefore may not have been coprecipitated with quartz (eg. Ebbutt 1948, Robert and Kelly 1987).

6.3.3.1. Canadian Archean

Oxygen isotopic data for vein quartz in Archean gold deposits in Canada have been reported by Kerrich and Fryer (1979), Kerrich and Hodder (1982), Fyon *et al.* (1983a), Fyfe and Kerrich (1984), Kerrich and Watson (1984), Kishida and Kerrich (1987), Brown and LaKind (1986), and McNeil and Kerrich (1986). Data from these studies are summarized on Figure 6.4. In all the Archean deposits, the $\delta^{18}\text{O}$ values of the vein quartz fall consistently in the range of +10 to +16 per mil. The $\delta^{18}\text{O}$ values for vein quartz are, for the most part, independent of host rock lithology. In addition, $\delta^{18}\text{O}$ values of vein quartz and immediately adjacent quartz in wall rock are the same; wall rock quartz has undergone significant positive changes relative to "fresh rock", towards equilibrium with vein quartz (eg. Kerrich and Fryer 1979, McNeil and Kerrich 1986). This indicates that the veins and their immediate wall rock environments were fluid-dominated, that the fluid represented a relatively high ^{18}O reservoir and that the fluid was introduced from a source external to the immediate depositional site. This suggests that, in the majority of cases, high fluid to rock ratios had been established throughout the mineralized zone. Temperatures calculated from mineral pairs (usually a combination of vein quartz, chlorite and muscovite) range from 320° to 480°C and estimates of fluid $\delta^{18}\text{O}$, in equilibrium with these phases, mostly fall within the range of +5 to +12 per mil (Figure 6.4).

6.3.3.2. Yilgarn Block - Western Australia

Golding and Wilson (1987) present a comprehensive compendium of oxygen isotopic data for deposits in Western Australia (Figure 6.4). The striking feature of these data is the range in $\delta^{18}\text{O}$ values for vein quartz (11 to 14 per mil), which corresponds with the range of data for Canadian Archean gold deposits (Figure 6.4). Also, $\delta^{18}\text{O}$ values of vein quartz are consistent over vertical distances of hundred's of metres. The hydrothermal fluid from which the quartz precipitated is inferred to have had a $\delta^{18}\text{O}$ value of +6 to +8 per mil. This is somewhat narrower than the calculated range for the Canadian Archean gold deposits (+5 to +12 per mil).

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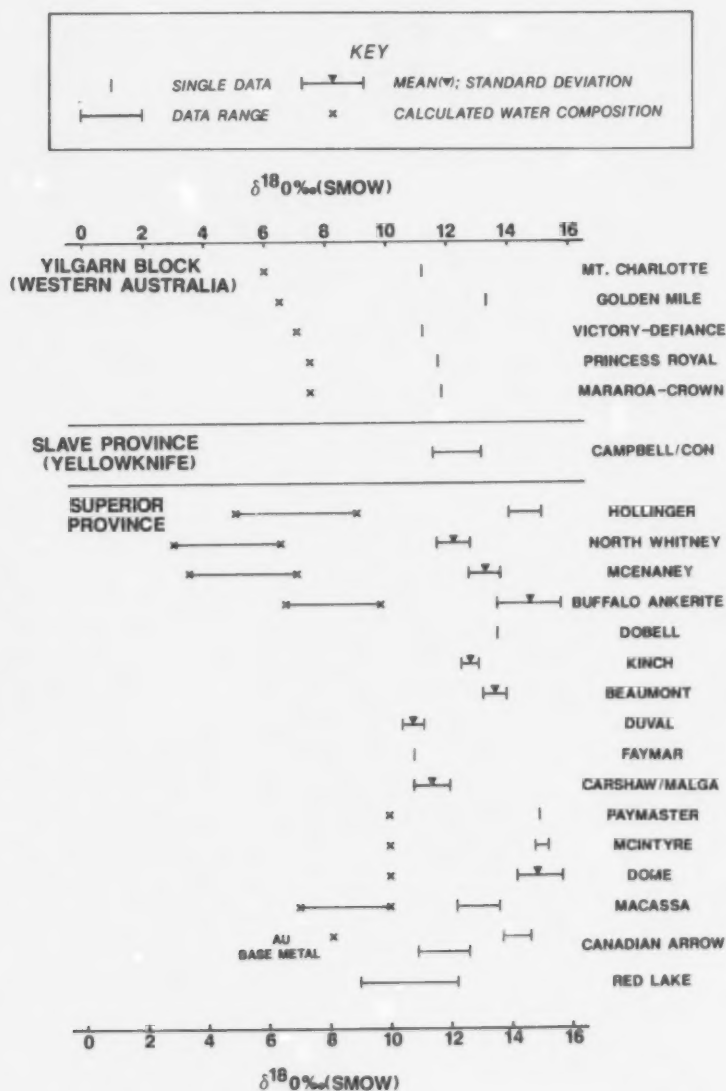


Figure 6.4. Partial compilation of $\delta^{18}\text{O}$ values for vein quartz in Archean gold deposits. Data taken from: Kerrich and Fryer (1979), Kerrich and Hodder (1982), Kerrich and Watson (1984), McNeil and Kerrich (1986), Brown and LaKind (1986), Fyon et al. (1983a), and Fyon (1986).

6.3.3. Summary

The calculated $\delta^{18}\text{O}$ of the hydrothermal fluids from which vein quartz precipitated ranges from +5 to +10 per mil. The vein system was fluid-dominated, and the wall rocks had equilibrated thermally with the fluid. Considering the vertical and lateral scale of some of the mineralized systems (eg. the Kirkland Lake system), the volume of fluid must have been large. Thus, the source of the fluids was external to the immediate depositional site. Temperatures within the mineralized zones

ranged from 300° to 500° C.

6.3.4. HYDROGEN ISOTOPIC DATA

Hydrogen isotopic data for Archean gold deposits are fewer than the sulphur, carbon or oxygen isotopic data. This partly reflects the difficulty in interpreting the data, and assessing the influence on silicate δD values of progressive metamorphic and hydrothermal fluid activity in and around the gold systems.

6.3.4.1. Abitibi Belt - Superior Province

Kerrick and Watson (1984) and Kerrich (1983) reported coincident hydrogen and oxygen isotopic data for lode gold deposits in the Macassa Mine at Kirkland Lake. These data fall within the range of -30 to -70 per mil. Hydrogen isotopic data for water extracted from fluid inclusions in quartz veins in the Timmins area were reported by Fyon *et al.* (1983a) and Fyon (1986). These data show a wide range, from approximately +10 per mil to -60 per mil, which may reflect partitioning between reduced and oxidized H-bearing fluid species due to changing redox conditions, or mixing of fluids from different sources (Fyon *et al.* 1983a).

6.3.4.2. Yilgarn Block - Western Australia

Golding and Wilson (1987) reported data collected from the Princess Royal and Victory deposits. Hydrothermal biotite has a δD value of -56 per mil. The δD values of water in equilibrium with this (hydrothermal) biotite in the Princess Royal deposit are -10 to -16 per mil, dependent on the deduced conditions of formation (metamorphic or hydrothermal). Chlorites from the Victory Mine (Kambalda) have a δD of -73 per mil (± 10 per mil; Golding and Wilson 1987). The calculated δD of water in equilibrium with these chlorites is approximately -30 per mil at 350°C (Golding and Wilson 1987).

6.4. SUMMARY

The striking similarity of fluid constitution and oxygen, hydrogen, carbon and sulphur isotopic data from Australian and Canadian Archean gold deposits supports the premise, based on the systematic patterns of alteration (see Chapter 3), that a compositionally uniform fluid was involved in the gold-related alteration and gold precipitation events, regardless of deposit location. This implies uniformity in the fluid generation process, involving similar types of rocks and reservoirs. The fluid was derived from a large reservoir, which was external to the immediate depositional environment. This reservoir appears to have been relatively uniform isotopically. Interpretations of these data as they pertain to specific aspects of this fluid reservoir, and the processes by which the fluids were generated, are considered in more detail in Part II. An interpretation of these data in terms of potential sources is also deferred to Part II (Chapter 1).

PART I. DEPOSITIONAL MODEL

7. DEPOSITIONAL MODEL

7.1. INTRODUCTION

In this report, a depositional model is distinguished from a genetic model in that it represents a synthesis of empirical data drawn from many Archean lode gold deposits, with a minimum of genetic interpretation. It is unlikely, however, that the inherent characteristics of these deposits would display systematic inter-relationships unless they were the product of a unifying genetic process. The only genetic process invoked in Part I is the introduction of large volumes of hydrothermal fluid of unique composition upwards into the depositional environment, necessitating a source external to the depositional environment. Evidence to support that premise was presented in Chapter 6.

Where the fluids and gold came from, and how they were transported are not essential components of a depositional model. Evidence is presented in Part II for a holistic genetic process of at least crustal scale. While evidence for such a process may support it, the depositional model is, for all practical purposes, independent of the genetic model. Hence, the paper is separated into Parts I and II.

Metallogenetic processes may be diachronous and, therefore, evidence of similar timing is not essential to a holistic model. Evidence is presented in Chapter 5 of Part I and in Part II that gold deposition over large areas of the Shield was correlative in age. This suggests that, not only were they formed by a common genetic process, but also that they were formed during a short lived, but major, metallogenetic epoch; in such an epoch, the degree of systematic inter-relationships of deposit characteristics is highest.

This chapter first discusses the inter-relationships of deposit characteristics, based principally on data from chapters 1 through 4. However, the authors recognize the complexity of gold systems, and some of the problems encountered in their study are also addressed. Lastly, as this paper is intended to be of practical use, some of the implications of the depositional model to exploration are discussed.

7.2. DEPOSITIONAL MODEL

Figure 7.1 presents a composite, idealized, Archean lode gold system; it is highly simplified in that only selected characteristics can be diagrammatically represented. It does not imply that the

complete system developed in any one location, but draws upon portions of gold systems which are well displayed in specific locations.

The diagram represents an original vertical section of the composite gold system. It is implied that it also represents the present-day orientations of deposits, allowing for some modification of their geometries by progressive shearing and, in some instances, tilting by rotational uplift. A vertical scale is not defined, but it can be inferred from the vertical range shown for individual deposits. A composite vertical range of greater than 5 km but less than 10 km is implied, depending on the thermal gradient during mineralization.

The depth of the deposits beneath the paleosurface cannot be defined. Evidence for lower confining pressures, in those deposits interpreted to represent the highest exposed level of the system, indicates that it was influenced by approach to surface. A minimum burial depth of 2 to 4 km is consistent with geobarometric fluid inclusion data. Present-day erosional surfaces expose all levels of the system, as indicated by the upper depth range of individual deposits.

The principal variable represented in Figure 7.1 is, therefore, burial depth beneath paleosurface, which is indicated by the metamorphic grades of the enclosing rocks. Chapter 3 discusses the evidence for the temporal association between gold mineralization and peak metamorphism. The ambient pressure and temperature during mineralization exert the strongest overall control on the products of the mineralizing process; this is manifest in variable styles of mineralization and alteration. At progressively higher levels, mineralization changes from shear-parallel without prominent veining, through systematic, dilatant vein arrays, to pervasive breccia-style mineralization. Alteration assemblages display equally prominent variations. Pyrite is the characteristic alteration sulphide in most deposits, but pyrrhotite predominates in deeper level deposits. Ankerite is ubiquitous in most gold camps, but is rare to absent in deeper level deposits. Silicate assemblages display equivalent systematic variations.

A strong host rock control tends to affect the deposit characteristics at a more local scale than overall depth zonation. It is impossible to represent the many effects of contrasting lithological competency in Figure 7.1. For example, more competent lithologies will host discrete vein arrays which may pinch out in adjacent, less competent units. As a

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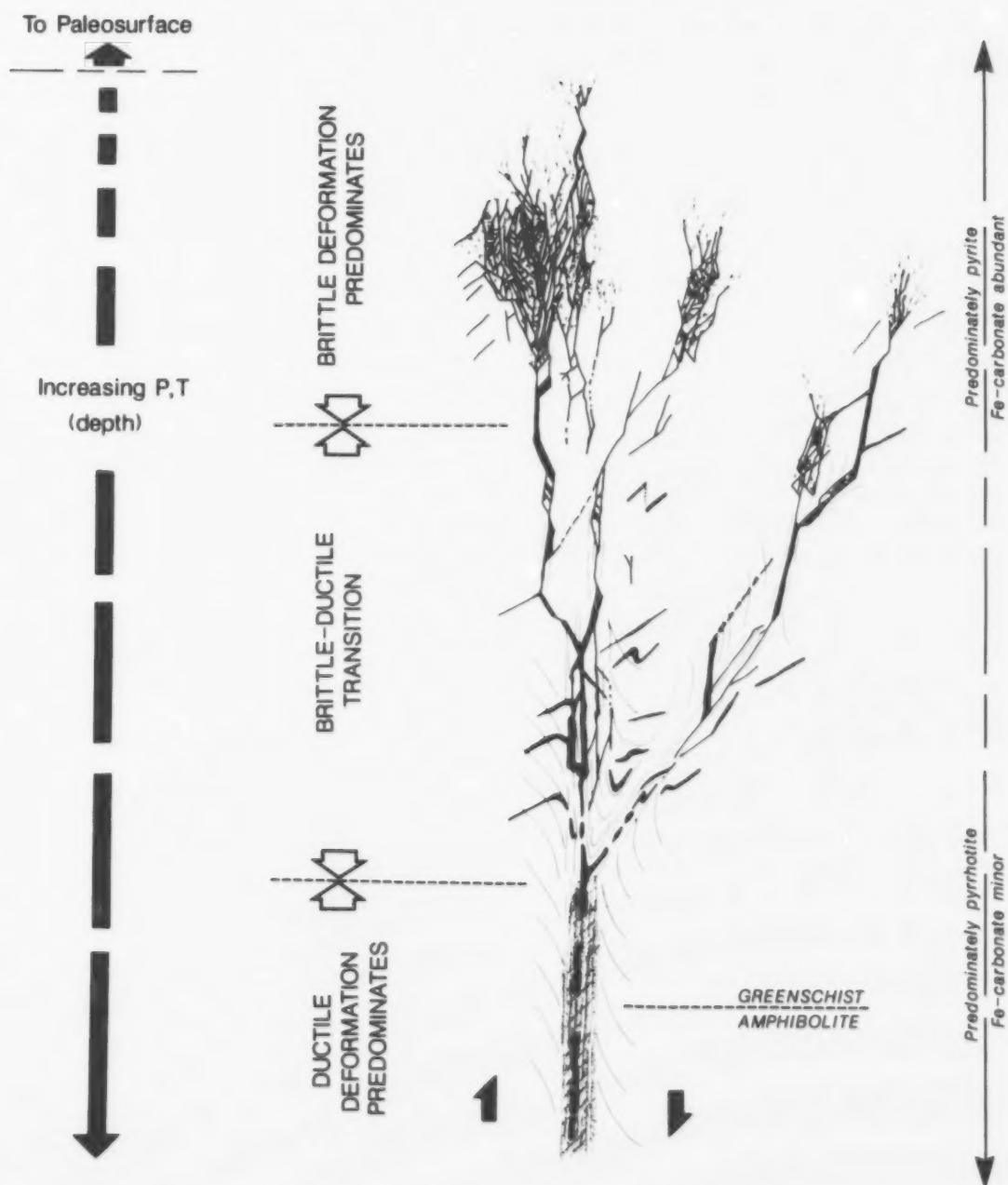
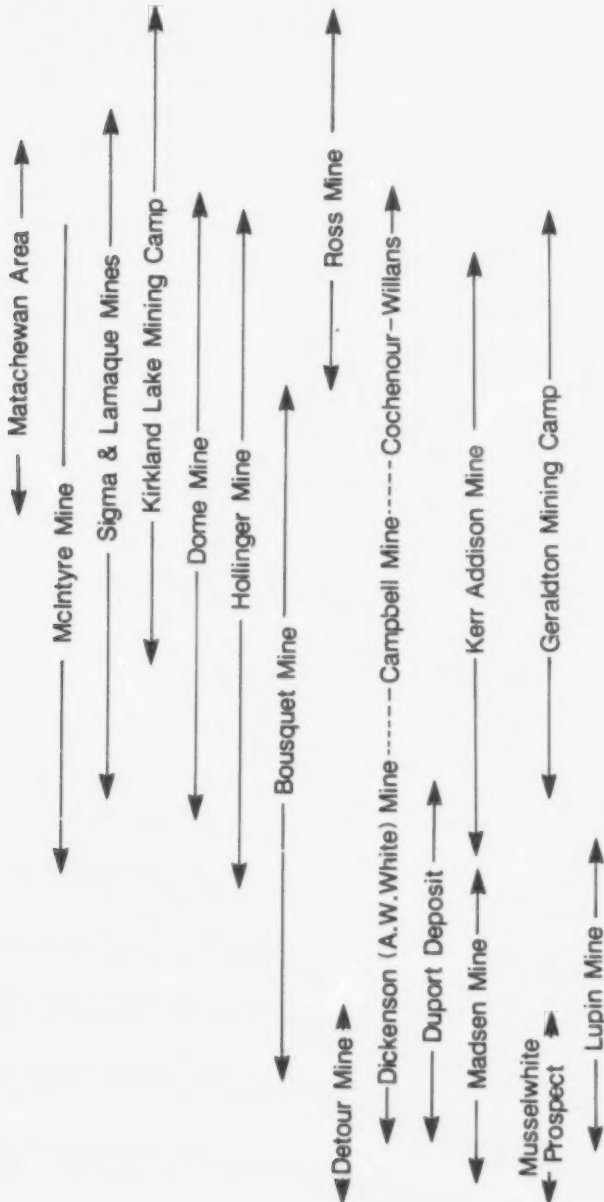


Figure 7.1. Idealized composite depositional model for Archean lode gold deposits.

...FELSIC.....BASALTIC KOMATIITIC..BIF..



Minerals common to upper level deposits

BASALTIC	BIF
White mica (+ Cr, etc.) Chlorite	Chlorite Stipnomelane
KOMATIITIC	FELSIC
Talc Tremolite Chrome mica Chlorite	White mica Albite + Hematite + Chlorite + Anhydrite + Potassium feldspar

+pyrite, +Fe-carbonate

Minerals common to lower level deposits

BASALTIC	BIF
Cordierite Staurolite Chloritoid Garnet Hornblende Biotite	Grunerite Hornblende Biotite Garnet
KOMATIITIC	FELSIC
Biotite Anthophyllite + Talc + Tremolite + Chrome mica + Chlorite	Potassium feldspar Phlogopite [Scapolite]

+pyrrhotite, -Fe-carbonate

PART I. DEPOSITIONAL MODEL

rule, competent units such as felsic intrusions retain a brittle response to deformation to deeper levels in the system, whereas less competent units, such as altered ultramafics, display a ductile response even at higher levels. Many deposit-specific examples of these controls have been discussed in the preceding chapters.

Alteration is also displayed on Figure 7.1, which shows the main alteration mineral phases in several lithologies at both upper and lower levels of the system. In most deposits, ore is hosted by several lithologies and it is important to recognize the mineralogical indications of equal alteration intensity in different rock types.

Hydrothermal fluids are an essential component of the gold system, and their effects can locally disrupt the systematic patterns displayed in Figure 7.1. For example, the addition of fluid can promote ductile response to shear, whereas fluid overpressure can result in local brittle response in a ductile regime. Fluid-rock interaction is inherent; carbonatization may increase rock competency whereas hydration, which produces phyllosilicates, may reduce it. Furthermore, pressure - temperature - fluid conditions in the mineralizing environment may vary considerably over time. The overall effects of the variability of the fluid in the gold system are, therefore, less easy to predict.

7.3. UNRESOLVED QUESTIONS

The above model attempts to summarize the effects of systematic, yet very complex, processes. Inevitably there are many aspects of these processes whose importance is difficult to establish. Their effects may be displayed as apparent exceptions to a general rule. Several of these have been identified by the authors and, no doubt, the reader will recognize others.

7.3.1. FELSIC INTRUSIONS

The common, but not ubiquitous, association between gold deposits and felsic intrusions has long been used as a prospecting guide. This spatial correlation, combined with the similar absolute age of these intrusions in Ontario, was considered highly significant by Colvine *et al.* (1984). However, those authors were undecided as to the role of these intrusions. They proposed two alternative hypotheses: that the intrusions were the source of the mineralizing fluid, or that they represented physical and chemical traps for mineralization. Evidence

presented in Chapter 5 of this report, and by Marmont and Corfu (1988), renders the former hypothesis of a "porphyry" model less likely in some deposits, although it may be valid in others (Smith 1987, Fyon and O'Donnell 1987). This has resulted in a re-assessment of the second alternative.

The favourable physical characteristics of the intrusions are more obvious. The intrusions are continuous over a large vertical range and retain a more brittle response to deformation at deeper levels; as such, they may represent the most favourable channelway for hydrothermal fluid ascent from depth.

Their chemical characteristics, particularly their relatively low iron content, ostensibly appear less favourable for gold precipitation. Sparse data indicate that some may have been highly oxidized at the magmatic-hydrothermal stage, as discussed in Chapter 4. Pre-existing, oxidized mineral species, such as sulphate and hematite, could react with later auriferous hydrothermal fluids, resulting in the precipitation of gold. Several large deposits are spatially associated with sulphate, including the Hollinger and McIntyre Mines at Timmins, the Kirkland Lake mines, the Hemlo camp, the Giant and Con Mines at Yellowknife, and the mines in Kalgoorlie, Australia.

The authors are not proposing that coincidence of these two processes is an essential component for gold concentration. Clearly, many gold deposits formed where there is no evidence for precursor oxidized assemblages. It is an intriguing hypothesis, however, which is consistent with the association between intrusions and some large gold deposits.

Regardless of their role, these intrusions are still an important component of many gold systems. Felsic intrusions are not diagrammatically represented on Figure 7.1, but are included on Figure 7.2 as an indication of the importance that the authors attribute to them. Further research into their geochemistry and petrogenesis is essential to better establish their role in mineralization, as they are now best characterized by geochronology.

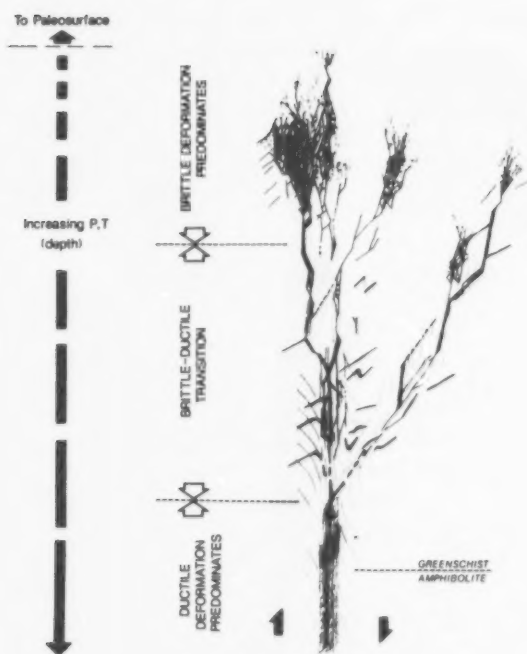


Figure 7.2. Idealized, composite gold system, from Figure 7.1, including a felsic intrusion.

7.3.2. METAMORPHISM

As discussed in Chapter 3, alteration took place penecontemporaneously with peak metamorphism. Alteration assemblages on Figure 7.1 reflect burial depths of peak metamorphism. At least three deposits do not fit this general rule:

(1) The Renco Mine in Zimbabwe is hosted by granulite facies rocks of the Limpopo Mobile Belt which have been overthrust onto the Zimbabwe craton from the southeast. Mineralization is hosted by brittle - ductile structures, which would not form in a granulite environment. Alteration assemblages are also consistent with those of lower amphibolite facies and clearly retrograded pre-existing granulite facies assemblages (Tabcart 1987). It is therefore suggested that mineralization took place substantially after attainment of peak metamorphism. This example opens the possibility of mineralization in terranes generally considered too high grade to host gold mineralization.

(2) At Hemlo there is conflicting evidence for the timing of mineralization relative to amphibolite grade metamorphism; both pre- and post metamorphic origins have been proposed (Kuhns *et al.* 1986, Burk *et al.* 1986). The reader is referred to the Hemlo section in the Gold '86 Proceedings Volume for comprehensive descriptions of the alteration and metamorphic minerals.

The Hemlo ore zones developed within a shear system; quartz veining is common, but the ore itself has been ductilely sheared. Pyrite is the predominant alteration sulphide and ankerite is rare. It is proposed that these apparently conflicting characteristics, some of which are indicative of a greenschist environment and others of an amphibolite environment, may have resulted from rapidly fluctuating thermal gradients during mineralization as a result of regional and contact metamorphism and fluid introduction.

(3) The Big Bell Mine in Australia is hosted by upper amphibolite grade rocks, in which partial melting has occurred (Phillips 1986, Phillips and DeNooy 1988). These authors interpreted this deposit to represent a highly metamorphosed gold deposit which originally formed in a greenschist environment. Absence of ankerite was attributed to decarbonation, although the expected calc-silicate assemblages are not reported. Pyrrhotite, which is the predominant alteration sulphide, was attributed to the desulphidation of pyrite. Retrograde alteration assemblages are reported but are not considered to be associated with mineralization (Phillips 1986). From the available literature, the described characteristics of the Big Bell deposit may also be consistent with mineralization after peak metamorphism, under ambient lower amphibolite conditions.

7.3.3. THE TOP OF THE SYSTEM

Present-day erosional surfaces expose sections of the latest Archean crust. The least erosion is represented by sub-greenschist assemblages, which may indicate original burial depths of about 2 km. Therefore, the upper level of exposed gold systems represents at least that burial depth. Although the indicated high fluid flux can best be explained by surface discharge, any products of that discharge

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would not be protected from erosion without subsequent infolding, volcanism or sedimentation. The question remains, however, whether the highest preserved level of the gold system have been discovered, or recognized.

Breccia zones on Figure 7.1 represent the highest identified level of the gold system within deposits in the Superior Province. Intuitively, these might be expected to change progressively into broader and pervasive mineralized and altered zones with decreasing confining pressures; boiling of fluids may also have occurred, as the paleosurface during mineralization is interpreted to have been subaerial. One possible example of this portion of the system is the Shamva Mine in Zimbabwe. Here, the "felsite" ore host constitutes pervasive K-feldspar alteration of quartz arenites or felsic volcanics. Pyrite and gold are somewhat irregularly distributed. At the deepest exposed level, about 400 m below surface, breccia-style mineralization, with high carbonate contents in the matrix, occurs. Alteration has obliterated any pre-existing structures and the ore zones are cut only by late faults.

If this interpretation of the Shamva deposit is correct, it opens up exciting exploration possibilities. The highly altered "felsite" could easily be mistaken for felsic volcanics, with a not unusually high pyrite content, yet the deposit may represent the upper end of a more extensive ore system. More than 2 million ounces of gold have been recovered from the upper 400 m of the Shamva Mine. Pyritic, siliceous-looking units in low metamorphic grade terranes, which actually represent albite or K-feldspar alteration, may be high potential exploration targets.

7.3.4. ORE MINERALOGY

As discussed in Chapter 3, it became obvious in the preparation of this paper that the available data on ore mineralogy and associated minerals were completely inadequate to assess their significance. While modern technology has rendered this research "old fashioned", it may provide critical information on ore paragenesis and have practical applications. This deficiency will be addressed by the authors, and hopefully by other researchers.

7.3.5. GOLD PRECIPITATION

How gold was transported in solution, and why it precipitated out of solution, remain somewhat speculative; there are little new data to help

resolve these questions. Thio complexing of gold is most commonly proposed as a method of transporting gold, but there is a lack of experimental data at high pressures and temperatures, and a need for additional solubility data between 250°C and 500°C (Seward 1984) to confirm this hypothesis. Without this information, explanations remain conjectural.

Gold deposition may be initiated by bulk oversaturation due to changes in temperature, pressure, Eh and Ph, or by varying activity of complexing ligands (Seward 1984). There appears to be no unique temperature or pressure of gold precipitation, although decreases in both were certainly contributory factors. In the Canadian Archean, evidence for gold deposition by fluid boiling has not been documented, although H₂O - CO₂ phase separation during hydraulic fracturing has been suggested by Spooner *et al.* (1987).

Destabilization of gold transported as a thio complex may also result from fluid oxidation. Among the reactions which may cause this are sulphidation of wall rocks, in which iron-bearing oxide and silicate minerals are replaced by iron sulphides (Macdonald 1984, Phillips and Groves 1984, Roberts 1987). Sudden oversaturation of gold may occur as a consequence of this sulphidation (Romberger 1986). Similarly, fluid reaction with precursor oxidized mineral species such as sulphate, as discussed above, presents an attractive alternative for rapid oversaturation of gold. However, evidence of spatially related oxidized mineral assemblages has been documented in only a few deposits, albeit that they are large.

Other, more local factors, may have caused gold precipitation. The piezoelectric properties of the alteration sulphides can cause rapid plating of gold onto sulphide minerals of the correct charge (Sakharova and Labachera 1978, Bancroft and Jean 1982, Mironov *et al.* 1981); this process may occur even if the fluid is undersaturated with respect to gold. This is consistent with the common occurrence of gold in alteration sulphides, rather than in quartz veins, as is well displayed at the Agnico Eagle Mine in Quebec (Barnett *et al.* 1982, Wyman *et al.* 1986). Similarly, gold is also intimately associated with carbon in some deposits (e.g. the Owl Creek Mine, Timmins). Shearing of carbonaceous sediments changes the structure of amorphous carbon to semi-graphite; the activated semi-graphite becomes a reductant and can be plated with gold. This is a natural analogue of the carbon-in-pulp

gold recovery process (Springer 1984, Wilson and Rucklidge 1986).

It is clear, therefore, that further experimental solubility and mineralogical data, which will help establish ore paragenesis, are required if the processes of gold precipitation are to be better understood.

7.4. EXPLORATION IMPLICATIONS

This model is based on data from many Archean lode gold deposits in Ontario. If valid, it may be used in exploration to predict characteristics of gold mineralization in a variety of geological settings.

Regionally, delineation of deformation zones is important. Within these deformation zones, more favourable target areas may be represented by extensional (pull-apart) environments, which can be identified by the widening of the deformation zones and the occurrence of Timiskaming-type sediments. At deeper erosional levels, some of the characteristics may not be present, or may not be prominent. Within the shear systems, both intersecting shears and felsic intrusions commonly represent the locus of gold concentration.

Figure 7.1 provides a degree of predictability to the expected characteristics of mineralization in many environments on a more local scale. For example, stratiform, foliation-parallel mineralization is more characteristic of areas of higher metamorphic grade, whereas cross-cutting vein and breccia mineralization predominates in lower metamorphic grade regimes. In heterogeneous lithological assemblages, mineralization may be focussed in the more competent units.

Alteration patterns can also be predicted on the basis of host lithologies and metamorphic environment. Of particular note is the absence of visually prominent carbonatization in higher metamorphic environments; the equivalent garnet-aluminosilicate assemblage is not so visually prominent and may easily be missed.

At the property scale, two characteristics may assist exploration of a specific occurrence. Firstly, virtually all ore zones are elongate and plunging; structural analysis commonly shows that ore zones plunge parallel to the stretch lineation, or to an intersection lineation. Intersection lineations may result from the intersection of two fabrics or

the intersection of a shear fabric and primary strata. Secondly, careful documentation of the mineralogy and alteration around any specific occurrence may also present a guide to ore, in drilling (e.g. Melling 1986).

Clearly, there are no simple answers to gold discovery, particularly as there are no universally applicable remote sensing techniques for gold. There are, however, predictive relationships of structural style, metamorphic grade and alteration mineralogy among other features. An understanding of the systematic inter-relationship of the geological characteristics of gold is therefore an important factor in enhancing the probability of success in exploration.

7.5. SUMMARY

An attempt has been made in Part I of this report to explain the variable characteristics of gold deposits on the basis of fundamental geological principles. The data base is predominantly empirical; additional empirical data which are consistent with this model do not constitute proof of the model, but do lend additional credence to it. Conversely, empirical data which are shown to be inconsistent with the model would put its validity into question. Scientific method and time are the test of any such hypothesis.

The authors realize that this paper is far from complete and that much work remains to be done. It was prepared at this stage because of an unwillingness to accept that "gold is where you find it"; it is hoped that it represents a basis for applying systematic geological studies in the important search for new mines in Ontario.

PART II. A GENETIC MODEL

1. FLUID SOURCES

1.1. INTRODUCTION

A summary of selected fluid inclusion and stable isotopic data for Archean gold deposits in Canada and Western Australia was presented in Chapter 6 of Part I. Possible sources of water and carbon in the assumed mineralizing fluid can be identified by comparing the fluid inclusion and stable isotopic characteristics of the Archean auriferous hydrothermal system with well characterized modern hydrothermal systems, and other fluid reservoirs which existed during the Archean.

1.2. FLUID INCLUSION CONSTRAINTS

Higher fluid pressures, and the absence of evidence documenting the boiling of water, suggest that the auriferous fluid evolved under lithostatic pressures of 1.5 to 4.5 kb (4 to 12 km; Ho 1987, Brown and Lamb 1986, Robert and Kelly, Wood *et al.* 1986, Smith *et al.* 1984). Fluid generation at a greater crustal depth would inhibit boiling; however, some Archean gold deposits display vein arrays which may have been generated by high fluid pressure under lower lithostatic pressure. However, fluid boiling can not be completely rejected without further data (eg. Ross Mine, Abitibi belt; Troop 1986). Thus, although the development of crackle breccia-type ore bodies is not as common as discrete vein ore bodies, some Archean gold systems may have been produced at crustal levels shallower than indicated by the fluid inclusion data base now available.

1.2.1. SALINITY

Although consistently low salinities of Au-bearing fluids are not unique to Archean gold deposits, direct comparison with other fluid systems must be made cautiously, because the role of immiscible fluid separation must be considered when comparing fluid salinities. Fluid inclusion filling temperatures and oxygen isotope geothermometry (see Chapter 6, Part I) indicate that gold precipitated in the temperature range of 350° to 500°C, at approximately 2 kb pressure. Under these conditions and given the deduced salt composition (1-6 wt %) of the initial, miscible Archean auriferous fluid, unmixing into CO₂- and saline H₂O-rich components is predicted (Bowers and Helgeson 1983). It is important to interpret the significance of fluid inclusion populations in this context. Evidence for immiscible separation of a dilute, CO₂-rich phase and a saline, H₂O-rich phase containing up to 35 wt. % NaCl equivalent is observed in several

Archean deposits (eg. Robert and Kelly 1987, Smith *et al.* 1986, Wood *et al.* 1986, Macdonald 1984, Ho 1987), and similar, coexisting fluid inclusion populations are reported from other deposits (eg. Brown and LaKind 1986). Thus, the salinity of some fluid inclusion populations (the saline, aqueous component) in Archean gold - quartz systems may approach that observed in Phanerozoic magmatic and meteoric fluids, some of which show evidence of boiling (eg. 35 wt % salt in porphyry Cu; 10 to 45 wt % salt in porphyry Sn - W, and >5 wt. % salt in epithermal Au - Ag; Spooner 1981, Hedenquist and Henley 1985). Therefore, salinity alone is not a particularly distinctive characteristic of Archean auriferous fluids.

1.2.2. COMPARISON WITH OTHER ORE FLUIDS

The combination of temperature, salinity and significant CO₂ content distinguishes Archean auriferous fluids from those associated with other types of mineral deposits. Diagenetic, saline fluids in sedimentary basins are characterized by low temperatures (80° to 150°C) and moderate to high salinities, but negligible CO₂ contents (Zimmerman and Kesler 1981). Fluids associated with volcanogenic massive sulphide mineralization have salinities similar to that of sea water (3.5 wt % dissolved solids), and temperatures in the range of 200° to 350°C, but low CO₂ contents (Spooner and Bray 1977). Epithermal precious metal deposits are dominated by meteoric fluids having low CO₂ and low salinity; these fluids may or may not have undergone boiling (Bodnar *et al.* 1985, Hedenquist and Henley 1985). Magmatic fluids related to mineralization associated with high-level granitic and granodioritic intrusive rocks (porphyry Cu - Mo, Sn, and W mineralization) are commonly ultrasaline (40-60 wt % NaCl equivalent) aqueous brines with variable CO₂ contents and were trapped at high temperatures (400° to 700°C; Nash 1976, Kelly and Rye 1979).

1.2.3. FLUID INCLUSION SUMMARY

Archean auriferous fluids are distinguished from basinal brines or high level, Tertiary intrusion-related hydrothermal systems by the collective characteristics of lower salinity, persistent CO₂, absence of H₂O boiling, higher trapping pressure, and moderate temperature. Fluids having these characteristics can be produced within low to medium grade metamorphic terranes. CO₂-rich fluids are typical of high grade (granulite facies) metamorphic terranes (eg. Crawford 1981, Touret and Dietvorst

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1983). Low salinity, CO₂-bearing (rich?) magmatic fluids can also evolve if CO₂ saturation is attained in an H₂O - CO₂-bearing silicate magma (eg. Holloway 1976). Thus, both metamorphic and magmatic process can produce fluids which are generally similar in composition to those implicated in the Archean gold event.

1.3. STABLE ISOTOPIC CONSTRAINTS

The stable isotope data for some Archean gold deposits were presented in Chapter 6, Part 1. This type of information, used in combination with geological and petrographic data, can provide useful information on thermal regimes of ore precipitation. In addition, permissive sources of the fluids and some of the constituent elements involved can be constrained by comparison with isotopically well characterized fluid reservoirs. In this section, the constraints on the sources of the water and carbon constituents of the auriferous hydrothermal fluid, provided by the stable isotopic data, are assessed.

1.3.1 OXYGEN AND HYDROGEN ISOTOPIC CONSTRAINTS

Calculated $\delta^{18}\text{O}$ and δD values for the water component of the ore fluids lie in the range +2.5 to +10.0 per mil and 0 to -70 per mil, respectively. The data lie in the overlapping fields of magmatic and metamorphic fluids (Figure 1.1). The additional constraints that the fluid was of uniform composition, regardless of deposit location, and was derived from a large reservoir of relatively uniform isotopic composition, external to the immediate depositional site. These facts are equally consistent with magmatic or metamorphic sources. The large range in δD values (0 to -60 per mil) may be evidence that redox conditions of the fluid changed at the depositional site, resulting in hydrogen isotope fractionation between H₂O and a reduced, H-bearing species (CH₄ or H₂; Fyon *et al.* 1983a, Fyon 1986).

Meteoric and sea water are important sources for hydrothermal fluids in modern regimes. Minor amounts of CO₂ are recorded in fluids from epithermal Au - Ag deposits, which are interpreted to have fluid systems dominated by meteoric water (e.g. Bodnar 1985). Nesbitt *et al.* (1986) suggest that low salinity, CO₂-rich fluids can be generated by deep meteoric circulation. However, the oxygen and hydrogen isotopic data for Archean gold deposits are not consistent with a meteoric fluid, unless the fluid had undergone extensive oxygen and hydrogen isotopic exchange with the rock so as to

become isotopically indistinguishable from either metamorphic or magmatic water.

Thus, the oxygen and hydrogen isotopic data are most consistent with either a magmatic or metamorphic source, but do not unequivocally discriminate between them.

1.3.2. CARBON ISOTOPIC CONSTRAINTS

Under certain circumstances, possible sources of carbon in the hydrothermal fluid can be inferred from the carbon isotopic value of a precipitated carbonate mineral. However, the $\delta^{13}\text{C}$ of carbonate minerals depends on the temperature, Eh, pH, and both the carbon isotopic value of and concentration of total dissolved carbon in the parent solution (eg. Ohmoto 1972); therefore, estimation of the carbon isotopic value of the parent hydrothermal solution, based on the $\delta^{13}\text{C}$ composition of a precipitated carbonate, must be carried out cautiously. The observed $\delta^{13}\text{C}$ values for carbon derived from various reservoirs which have been positively identified in or can be reasonably inferred to have existed during the late Archean are presented in Figure 1.2. The calculated $\delta^{13}\text{C}$ value for the total dissolved carbon in the Archean auriferous fluid is approximated by that value for the gold-related carbonates (-8 to -3 per mil; see Chapter 6, Part I).

1.3.2.1. Magmatic Carbon

The inferred $\delta^{13}\text{C}$ for the hydrothermal carbon (-8 to -3 per mil) falls within the range observed for carbon from diamonds, kimberlites and carbonatites (-2 to -12 per mil; Koval'skiy and Cherskiy 1973, Kobelski *et al.* 1979, Sheppard and Dawson 1973, Deines and Gold 1973), and is consistent with a magmatic source for this carbon. The median $\delta^{13}\text{C}$ value for the Archean replacement and gold-related carbonates is about -3.5 per mil (see Figure 6.1 in Part I), which lies at the upper end of the range for magmatic carbon (Figure 1.2). However, some modern geothermal systems release CO₂ gas, interpreted to be of magmatic origin, which has an average $\delta^{13}\text{C}$ of -3 per mil (summarized by Hoefs 1980). The CO₂ in these geothermal systems coexists with a few mole percent CH₄ having negative δD values (eg. -25 per mil; summarized in Hoefs 1980). Carbon isotope fractionation at 200° to 300°C, during the formation of about 5 mole percent CH₄, at the expense of CO₂ and H₂O in the fluid, shifted the $\delta^{13}\text{C}$ of the residual CO₂ from an initial value near -5 per mil, to -3 per mil. Such modified magmatic CO₂ gas

has a $\delta^{13}\text{C}$ value which approximates the median $\delta^{13}\text{C}$ value for Archean carbonates from the regional and auriferous deformation zones. Minor amounts of CH_4 are also present in the Archean systems (eg. Wood *et al.* 1986, Robert and Kelly 1987), although the $\delta^{13}\text{C}$ of this methane is not known.

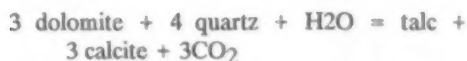
Thus, by analogy with some modern geothermal systems, estimates of the $\delta^{13}\text{C}$ of CO_2 in the Archean auriferous system may be too high, because of carbon isotope fractionation between oxidized and reduced carbon species in solution. Given that the gold-related fluids contained only a few mole percent CH_4 (Wood *et al.* 1986, Robert and Kelly 1987), and assuming that this methane formed in the fluid in response to changes in redox conditions, then any coupled carbon isotopic shift in the coexisting CO_2 would also be small, but significant. Thus, the "best" estimate for $\delta^{13}\text{C}$ of the total dissolved carbon in the hydrothermal fluid, based on the $\delta^{13}\text{C}$ values of the carbonate minerals (-3.5 per mil), serves as an approximate upper limit. The actual value was probably smaller.

To assess the $\delta^{13}\text{C}$ of the total dissolved carbon in the gold-related hydrothermal fluid, Fyon (1986) extracted this carbon by decrepitation and oxidized the gas with CuO . The $\delta^{13}\text{C}$ of this total dissolved carbon in fluid inclusions from the Carshaw/Malga (Gail Resources Ltd.) iron formation-hosted gold deposit near Timmins is about -7 per mil. This value is identical to that of oxidized magmatic carbon (eg. Pineau *et al.* 1976). Thus, the $\delta^{13}\text{C}$ of the gold-related hydrothermal fluid may be closer to the accepted magmatic composition (-5 to -7 per mil) than is suggested in the literature.

1.3.2.2. Metamorphic Carbon

CO_2 gas with $\delta^{13}\text{C}$ values similar to the Archean gold-related carbonates can be produced by several metamorphic processes when carbon- and carbonate-bearing rocks undergo burial and heating:

1) decarbonatization:



2) hydrolysis: $2\text{C} + 2\text{H}_2\text{O} = \text{CO}_2 + \text{CH}_4$

3) oxidation: $\text{C} + \text{O}_2 = \text{CO}_2$

4) dissolution: $\text{CaCO}_3 + 2\text{H}^+ = \text{H}_2\text{CO}_3 + \text{Ca}^{2+}$

For a typical Archean terrane, four carbonate reservoirs could yield CO_2 gas upon burial and metamorphism: 1) dispersed calcite; 2) sedimentary carbonaceous units; 3) marine carbonate sediments, and 4) recycled calcite and dolomite from zones of carbonatization.

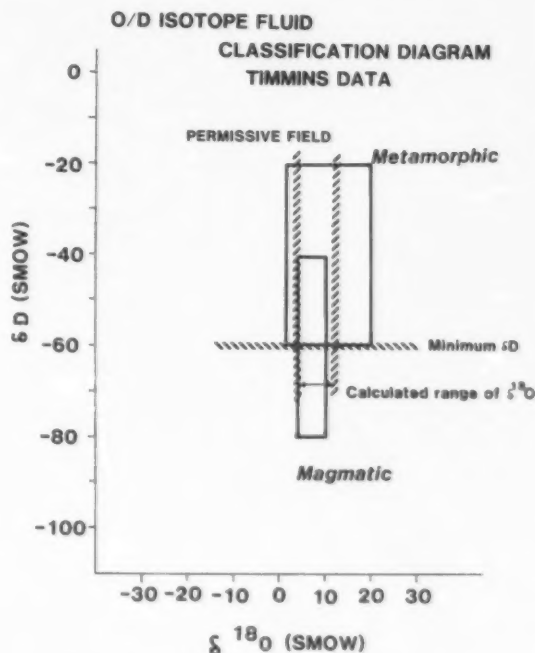


Figure 1.1. $\delta^{18}\text{O}$ vs. δD isotopic fluid classification diagram, showing the empirical fields for magmatic and metamorphic fluids, and the meteoric water line. The fields for data from Archean gold deposits are indicated.

Decarbonatization and Dissolution

When carbonate minerals undergo partial decarbonatization, the CO_2 gas produced has a $\delta^{13}\text{C}$ value which is 3 to 5 per mil heavier than the original carbonate (Shieh and Taylor 1969, Deines and Gold 1969, Sheppard and Schwarz 1970, Schidlowski *et al.* 1979). Decarbonatization of marine carbonate (dispersed calcite and Archean carbonate sediments; $\delta^{13}\text{C}$ values ranging from +3 to -3 per mil, mean 0 per mil, see Figure 6.1 in Part I) would have yielded CO_2 gas having a $\delta^{13}\text{C}$ value of 0 to +5 per mil, higher than the value calculated for the hydrothermal fluids. Dissolution of carbonate sediments may also yield CO_2 -bearing fluids having a $\delta^{13}\text{C}$ comparable to that of the protolith (+3 to -4 per mil). Assuming complete dissolution, these sediments would yield a CO_2 -bearing fluid having a $\delta^{13}\text{C}$ averaging 0 per mil.

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The relatively small volume of Archean sedimentary carbonate material restricts the generation of large volumes of CO₂-bearing hydrothermal fluid. Furthermore, carbonate sediments are more common in Archean platformal sequences, which are not abundant in terranes such as the Abitibi belt. Thus, the distribution of gold mineralization is clearly not dependent upon the presence of carbonate sediments. Therefore, while isotopically permissive, dissolution of marine carbonate is not considered to be a dominant mechanism in the generation of the gold-related fluid.

Decarbonization or dissolution of the ferroan carbonate present in the intense carbonate-altered zones could yield CO₂ having a $\delta^{13}\text{C}$ ranging from -5 to +3 per mil, based on the observed range of $\delta^{13}\text{C}$ for the replacement carbonate (-8 to 0 per mil; Figure 6.1 of Part I). Assuming a median value of approximately -4 per mil, the $\delta^{13}\text{C}$ of CO₂ gas produced this way would be too high, with respect to that deduced for the hydrothermal fluid. Also, despite variations in metamorphic grade, Golding *et al.* (1987) found no evidence for coupled ¹³C and ¹⁸O depletion in the carbon isotopic data from Western Australia, suggesting that little or no CO₂ was derived by decarbonization from this potential source. Hence, any metamorphic CO₂ contribution to the ore fluids, by recycling carbon from the regional zones of carbonatization, must have been derived by dissolution, which would yield produced CO₂ that is isotopically similar to the original carbonate.

Hydrolysis and Oxidation

Reduced carbon exists in Archean greenstone belts as carbonaceous sediments, and this material may have been a source of hydrothermal carbon through hydrolysis and oxidation at high temperature metamorphic conditions. Ohmoto and Rye (1979) suggest that the resultant CO₂ would probably have $\delta^{13}\text{C}$ values less than -10 per mil. Thus, it is unlikely that this material contributed much CO₂ to the auriferous hydrothermal fluid.

1.3.2.3. Isotopic Provinciality

Golding *et al.* (1987), Perring *et al.* (1987), Kerrich (1986) and Kerrich *et al.* (1987) emphasize the "pronounced provinciality" of carbon isotopic data and suggest that this favours the generation of CO₂ by metamorphic (greenstone belt - scale), rather than magmatic (mantle - scale) processes. However, the summary of the carbon isotopic data for mantle-derived carbon (Figure 1.2, Part II)

illustrates that the $\delta^{13}\text{C}$ values for mantle-derived CO₂ are not uniform (+0.2 to -33 per mil, at the extremes), although $\delta^{13}\text{C}$ of -5 to -7 per mil may be regarded as a reasonable approximation of the carbon isotopic value of magmatic CO₂ (eg. Pineau *et al.* 1976). However, this "provinciality" is typical of carbonatite carbon isotopic data (eg. Deines and Gold 1973). Furthermore, it is difficult to assess the degree to which the "provinciality" in the Archean data reflects differences in isotopic composition of the source, isotope fractionation between oxidized and reduced carbon species during fluid evolution from source up to and at the depositional site, or isotopic fractionation during progressive fluid/rock interaction at the depositional site (eg. see Chapter 6, Part I). Thus, while a component of the C-isotopic "provinciality" may reflect "greenstone belt - scale" processes, such as interaction of fluids with wall rock, carbon isotopic "provinciality" may be a general characteristic of the mantle (Deines and Gold 1973), and therefore, can not be used as evidence for or against magmatic or metamorphic sources.

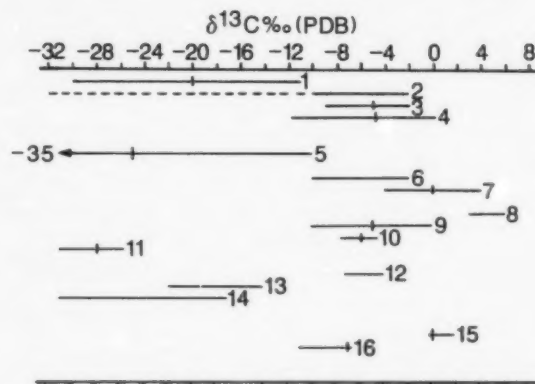


Figure 1.2. $\delta^{13}\text{C}$ values of terrestrial carbon material and sources. Data sources: Reduced carbon in igneous rock - Hoefs (1973), Fuex and Baker (1973), Pineau *et al.* (1976), Craig (1973); Diamonds - Koval'skiy and Cherskiy (1973), Kobelski *et al.* (1979); Carbonatites - Sheppard and Dawson (1973), Deines and Gold (1973); Kimberlites - Kobelski *et al.* (1979); CO₂ gas in vesicles - Pineau *et al.* (1976), Moore *et al.* (1977); Reduced C in organic material - Schwarcz (1969), Ohmoto and Rye (1979); Sedimentary carbonate - Veizer and Hoefs (1976), Schidlowski *et al.* (1975); Pure CO₂ in granulites - Hoefs (1975); CO₂ and H₂O in granulites (possibly retrograde) - Hoefs (1975); CO₂ in mantle nodules - Hoefs (1975); Modern atmosphere - Keeling (1958).

1.3.2.4. Summary of Carbon Isotopic Interpretation

The deduced $\delta^{13}\text{C}$ for the CO_2 from which the Archean, gold-related hydrothermal carbonates precipitated (-8 to -3 per mil) is most consistent with a magmatic reservoir (Burrows *et al.* 1986, Spooner *et al.* 1985, Fyon *et al.* 1983a), although a metamorphic origin, specifically dissolution of carbonate related to regional zones of carbonatization and regional faults (Perring *et al.* 1987), can not be ruled out on the basis of carbon isotopic data. However, field evidence and geochronological data demonstrate that all of this carbonate in the Superior Province was introduced coevally with the concentration of gold, into late regional structures, following at least one regional metamorphic and deformation event in the greenstone belts. Thus, the relative timing of alteration imposes serious constraints on this source of carbon, notwithstanding the compatible carbon isotopic values.

In the Timmins camp, the fluid may have experienced minor changes in redox conditions at the depositional site, which resulted in the precipitation of ^{13}C -enriched carbonate (see Chapter 6, Part I). The extent to which the fluids in other gold camps underwent similar changes in redox conditions at the depositional site is not known, but redox changes may account for at least some of the carbon isotopic "provinciality". In addition, redox changes may have contributed to the precipitation of gold (Chapter 5, Part II).

1.3.3. SULPHUR ISOTOPIC CONSTRAINTS

Like carbon, it is difficult to unambiguously characterize the sulphur isotopic values of all potential terrestrial sulphur reservoirs, because small changes in temperature, pH, and $f\text{O}_2$ can induce profound changes in the sulphur isotopic values of indigenous or derived sulphur-bearing species (Sakai 1968, Ohmoto and Rye 1979). Significant fractionations can potentially occur at the fluid/sulphur source, during fluid migration and transport, and at the site of precipitation during rock fluid interaction.

Given the specific environment of Archean gold mineralization, the most likely sources of sulphur, in terms of size and availability of reservoirs, include direct derivation from magmatic fluids and leaching of primary and secondary sulphide minerals from igneous rocks. Sedimentary marine sulphate and sea water sulphate are less plausible source reservoirs, given the rarity of unequivocal Archean marine sulphate deposits, which would be so small as to be insignificant in the present context. Fur-

thermore, sea water sulphate deposits are characteristic of the Archean platformal sequences. Gold deposits are much more widely distributed, and do not show a spatial correlation with either the evaporitic sediments or the terranes in which these sediments occur. Regardless, the $\delta^{34}\text{S}$ values of Archean sulphate deposits cluster near 0 per mil, indistinguishable from magmatic sulphur (Lambert *et al.* 1978). This leads to the inference that derivation of sulphur in Archean gold systems involved participation of either magmatic fluids and/or metamorphic dehydration/degassing of the underlying rock column.

1.3.3.1. General Isotopic Considerations

The following general points are relevant to the interpretation of the $\delta^{34}\text{S}$ data: 1) fluids having variable $\text{SO}_2/\text{H}_2\text{S}$ ratios would precipitate sulphides having a variable $\delta^{34}\text{S}$ (Ohmoto and Rye 1979); 2) sulphide minerals precipitated from a fluid whose dissolved sulphur was dominated by reduced species, without significant oxidation at the site of precipitation, would exhibit $\delta^{34}\text{S}$ values slightly more positive than that of the fluid and would have a small isotopic variation (Ohmoto and Rye 1979); 3) Fe-sulphide minerals generated from oxidized fluids, or from fluids undergoing significant oxidation at the precipitation site, would exhibit $\delta^{34}\text{S}$ values considerably more negative than that of the fluid; 4) sulphide minerals precipitated from a fluid that acquired its sulphur by leaching pre-existing sulphur-bearing minerals should have a small isotopic variation, but could have a $\delta^{34}\text{S}$ value higher (up to 5 per mil) than that of the leached, pre-existing mineral (Grinenko and Grinenko (1972).

1.3.3.2. Interpretation of Data

The extreme variation in $\delta^{34}\text{S}$ values for gold-related pyrite, which allows the subdivision of data into two broad groups (Figure 6.3) is perhaps the most distinctive feature of the sulphur isotopic data. Oxidized fluids are implicated in those deposits having negative $\delta^{34}\text{S}$ values for pyrite (Schwarcz and Rees 1985, Cameron and Hattori 1987); hence, the $\delta^{34}\text{S}$ of the total dissolved sulphur in the fluid should be larger than that of the precipitated pyrite (point 3, above). For those deposits having pyrite with $\delta^{34}\text{S}$ values of 0 or larger, fluids dominated by reduced sulphur species with only limited oxidation accompanying precipitation are implicated (Golding and Wilson 1983, Spooner *et al.* 1985); hence, the $\delta^{34}\text{S}$ of the total dissolved sulphur in the fluid should be lower than that of the pyrite (point 2, above). Lambert *et al.*

(1984) attribute the negative values obtained in the Golden Mile deposits to significant oxidation of the fluids during extensive reaction with rocks at the site of deposition. While applicable at an individual deposit, it seems that the general application of this premise may not be warranted, because the existence of hematite and sulphate minerals in many deposits implies oxidizing conditions (Cameron and Hattori 1987). The generation and evolution of oxidized fluids is examined in more detail in the following chapter.

1.3.3.3. Sulphur Isotopic Summary

Negative $\delta^{34}\text{S}$ values found in some Archean gold deposits can be attributed to physico-chemical conditions of the hydrothermal system, with changes in $f\text{O}_2$ being most significant (Cameron and Hattori 1987). A $\delta^{34}\text{S}$ of about 0 per mil for the total dissolved sulphur in the auriferous fluid is consistent with, and capable of yielding, the observed $\delta^{34}\text{S}$ values observed in Archean gold deposits, given appropriate reducing or oxidizing conditions at the depositional site. This isotopic composition may be interpreted as suggesting the parental fluids contained magmatic sulphur. However, this magmatic sulphur may have entered the fluid phase either by dissolution of juvenile sulphide minerals by metamorphic fluids, or by direct contribution from magmatic sources. Neither process is discriminated unequivocally by the sulphur isotopic data.

1.4. SUMMARY OF CONSTRAINTS

The collective consideration of the isotopic and compositional characteristics of the auriferous fluids and some related hydrothermal minerals identifies two reservoirs and attendant processes which could have produced fluids whose composition and isotopic characteristics would compare favourably with those of the high density, low salinity, CO_2 -bearing fluids observed in Archean lode gold deposits: magmatism and metamorphism (Macdonald 1984, Golding and Wilson 1987, Ho 1987, Perring *et al.* 1987, Kerrich *et al.* 1987, Burrows *et al.* 1986, Burrows and Spooner 1986; Groves and Phillips 1987). Neither the fluid inclusion nor the stable isotopic data can unequivocally distinguish between these two processes or fluids derived from them. Both processes are capable of providing large volumes of fluid, generated external to the depositional site.

In the following chapters, the process by which fluids are generated and evolve in both magmatic and metamorphic regimes are addressed. This exercise provides a means to refine the fluid generation models, and, thus, allows a more rigorous evaluation of their relative roles to the generation of the auriferous fluid.

2. MAGMATIC FLUIDS

2.1. INTRODUCTION

Fluid inclusion and stable isotope characteristics of Archean gold-related fluids suggests development by either a magmatic or metamorphic process. Because these techniques do not provide an unequivocal distinction, each process is examined in more detail, to identify possible fluid evolutionary paths. In this chapter, magmatic fluid generation is considered.

Classical, magmato-hydrothermal models of Archean gold mineralization have been revived (eg. Mason and Melnik 1986, Wood *et al.* 1984, 1986, Spooner *et al.* 1985, Burrows *et al.* 1986, Macdonald, 1984), partly because a plutonic association is characteristic of the gold depositional environment (Chapter 4, Part I), and because fluid inclusion and stable isotopic constraints are consistent with a magmatic fluid origin (Chapter 1, Part II). However, to critically assess the magmatic origin hypothesis, two important features of Archean gold mineralization and related fluids must be addressed:

- (1) Is it possible for a classical model of magmatic fluid generation, involving silicate magma generation, fluid exsolution, and mineral precipitation, to generate fluids compatible with those attributed to the Archean gold-related fluids?

- (2) It has been demonstrated that the emplacement of intrusive bodies within and adjacent to regionally extensive zones of deformation, late in the tectonic history (Chapters 2 and 4, Part I), was characteristic of Late Archean greenstone belt development. Yet, syn-volcanic plutonism was also a prominent characteristic of greenstone belt development, manifest as tonalite-trondhjemite intrusions, consanguineous with volcanism (eg. Paradis *et al.* 1988). Despite this long history of plutonism in parts of the Superior Province, significant gold mineralization developed only during late stages of tectonism and plutonism, after the termination of calc-alkaline volcanism and consanguineous plutonism (Chapter 5, Part I).

If gold introduction is attributed solely to magmatic processes, there must have been something "unique" about the process by which the late plutons were generated, their source, magma type and/or their magmatic fluid evolution. Some of these aspects are addressed cursorily in the follow-

ing sections, to suggest which may be important and which cannot be addressed adequately, because of insufficient data. The models of fluid evolution discussed in this chapter are not intended to be comprehensive. The discussions center on the relevance to Archean gold metallogeny. Other comprehensive discussions of magmatic fluid evolution are available elsewhere (eg. Burnham 1979).

2.2. MAGMATIC FLUID EXSOLUTION

Magmatic fluids are produced when a silicate magma attains vapour saturation and a vapour phase exsolves from the silicate melt and crystal system. The presence of CO₂ is a fundamental characteristic of the Archean auriferous, hydrothermal fluids, but CO₂, while being a conspicuous component, does not appear to be a dominant constituent in Phanerozoic, magmatic, ore-forming fluids. For example, porphyry Cu fluids generally contain less than 3 mole % CO₂, although concentrations up to 30 mole % have been reported (Nash 1976). Were Archean magmatic fluid essentially identical to those of modern intrusions? Do differences exist in the constitution of magmatic fluids which can be related to differing source conditions? In the following sections, the roles of differing source compositions, and CO₂ and salt concentrations, on the constitution and evolution of magmatic fluids are examined.

2.2.1. EFFECT OF CO₂ ON VAPOUR SATURATION

CO₂-rich, aqueous magmatic fluids can be generated at various crustal levels by "normal" crystallization of a tonalite-granite magmas, if the magma source area contained both H₂O and CO₂. CO₂ solubility in silicate melts is dependent on melt composition, temperature and pressure, and is lower than that of H₂O (Figure 2.1; eg. Eggler 1974, Kadik and Eggler 1975, Mysen 1976). Because of the low solubility of CO₂ in the magma, silicate melts need to contain only small amounts of volatiles to become CO₂-saturated. Since CO₂ solubility in the magma is greater at high pressure and temperature (Figure 2.1), melts generated under mantle conditions and then emplaced into the crust will become saturated with CO₂-rich vapour phases before H₂O saturation is attained (Holloway 1976). Therefore, the melt phase will coexist with a vapour phase which is richer in CO₂ (Figure 2.2; eg. Wells 1979). Thus, vapour-saturated melts will evolve CO₂ (plus minor H₂O) vapour continuously during ascent from their source region, and will

PART II. GENETIC MODEL

crystallize at higher temperatures than magmas that are H₂O-saturated if the CO₂ vapour remains in equilibrium with melt and crystals.

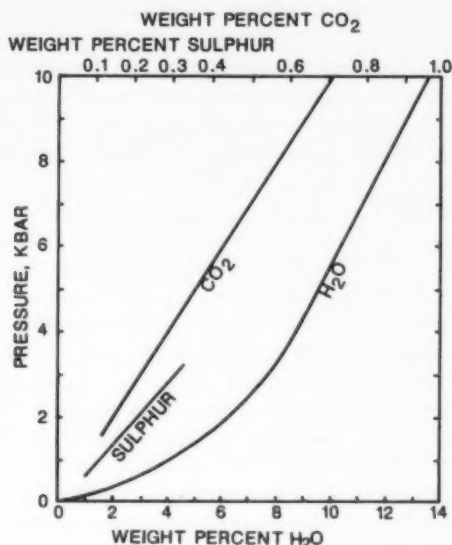


Figure 2.1. Solubility of CO₂, H₂O and S in a "felsic" melt as a function of temperature and pressure. Sulphur data from Carroll and Rutherford (1985), solubility for dacite melt at 1025°C and fO_2 of hematite-magnetite buffer. Data for CO₂ from Holloway (1976), solubility for binary CO₂ - H₂O fluids in albite liquid, $XCO_2(\text{fluid}) = 0.75$. Data for H₂O from Burnham and Ohmoto (1980). Diagram from Cameron and Hattori (1987).

If vapour is progressively distilled during near-isothermal ascent, the magma will remain molten to lower temperatures. The water content of the melt will increase, but the vapour phase in equilibrium with melt and crystals will become enriched in CO₂ until pressure is reduced to about 5 kb. At about 4 to 7 kb, the CO₂:H₂O ratio in the vapour phase distilled from the magma reaches a maximum (eg. Holloway and Lewis 1974, Wells 1979). At lower pressure, the CO₂:H₂O ratio in the evolved vapour is decreased (Figure 2.2). If primary crystallization of the tonalitic melt takes place at pressures greater than 4 to 7 kb, most of the H₂O dissolved in the melt could be taken up by crystallizing silicates, such as biotite and hornblende (Stern *et al.* 1975), and the final composition of the evolved vapour phase could become very CO₂-rich.

Thus, for a silicate magma to become

vapour saturated at a higher temperature early in its ascent, and to evolve CO₂-rich vapour, it is essential that the anatexic source area contain some CO₂. In the absence of CO₂ in the source area, vapour saturation may be achieved at a higher crustal level, at lower temperatures, and the vapour could be dominated by H₂O, with only traces of CO₂.

2.2.2. EFFECT OF DISSOLVED SALTS

Exsolved magmatic fluids contain NaCl. If present in sufficient quantities in an H₂O - CO₂ fluid, NaCl will induce the unmixing of the CO₂ - H₂O vapour to yield a CO₂-rich component and an aqueous brine component, even at temperatures in excess of 1000°C (Bowers and Helgeson 1983, Trommsdorff and Skippen 1986). Fluid unmixing would further enrich the CO₂ content of the vapour, which, having a density different from that of the saline component (Trommsdorff and Skippen 1986), could infiltrate the country rocks, away from the silicate melt and the companion saline aqueous fluid. The coexistence of saline, aqueous fluid inclusions with CO₂-rich, low salinity fluid inclusions, observed in many gold deposits, has been attributed to fluid unmixing in the depositional site (eg. Robert and Kelly 1987, Wood *et al.* 1986). However, fluid unmixing could take place at crustal depths and temperatures significantly below that of the gold depositional site, dependent on the initial CO₂ and dissolved salt content.

Thus, the combined effect of dissolved salt and CO₂ could produce a complicated magmatic fluid evolution. A spectrum of fluid compositions could result, dependent on the depth at which exsolution took place, the degree to which the exsolved fluid remained in equilibrium with the silicate magma, and whether the fluid underwent immiscible separation. In many late zones of deformation, carbonate occurs in two distinct associations - nearly pure dolomite - ankerite veins, with little or no quartz, and as an accessory mineral in quartz-dominated veins. The presence of two modes of carbonate occurrence within the same structural setting, but at different times (see Chapters 3 and 6, Part I), implies that the attendant CO₂-bearing fluids may have followed quite different evolutionary paths, compatible with the general principals discussed above.

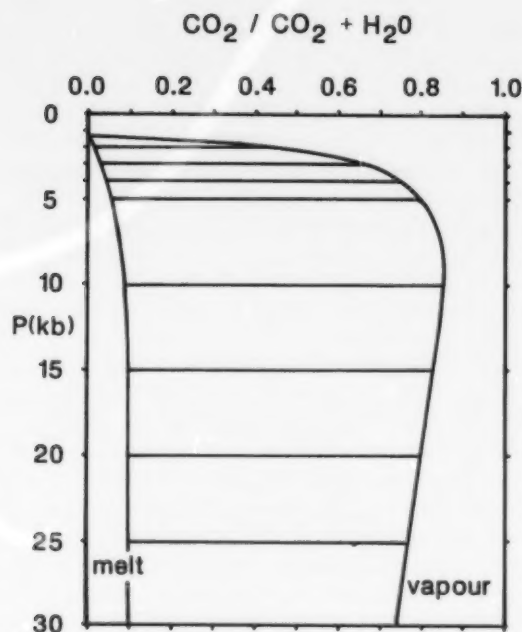


Figure 2.2. Estimated $P - X(\text{CO}_2)$ diagram showing phase relations and compositions of tonalitic liquid and vapour phase coexisting with crystals at an approximate temperature of 950°C . Data base from Egger (1974), Egger and Burnham (1973), Kadik and Egger (1975), Hill and Boettcher (1970) and Millhollen *et al.* (1971). Diagram from Wells (1979).

2.3. LATE PLUTON - GOLD ASSOCIATION

The spatial and temporal association between Archean gold mineralization and plutonic rocks was discussed in Chapter 4, Part I. Possible differences between plutons which were emplaced late in the magmatic history of greenstone belts and those which were essentially syn-volcanic are considered in this section, to establish if one or more characteristics exist which could account for the frequent spatial and temporal association between gold and the late plutonic group.

2.3.1. MAGMA COMPOSITION

A significant difference between the syn-volcanic intrusions and the late intrusions is the compositional variety represented in the latter (Chapter 4, Part I). Virtually all of the syn-volcanic, "felsic" intrusions are members of the tonalite - trondhjemite - granite (TTG) suite (eg. Corfu and Andrews 1987, Corfu and Stott 1986). Intrusions

belonging to the TTG suite were also emplaced during the late plutonic event; however, intrusive complexes having varying degrees of silica saturation are restricted to the late plutonic event, based on the present field, petrological and geochronological data bases. Thus, differences in magma type and, therefore, source and genetic process are implied by the compositional diversity of magmas in the late plutonic group, relative to the "syn-volcanic" plutonic group. This may indicate direct mantle involvement during the late plutonic event (eg. Shirey and Hanson 1984).

The conspicuous presence of silica-under-saturated magmas in the late plutonic suite, sometimes spatially associated with gold, lends support to the premise that this magma type may be genetically involved in the Archean gold metallogenesis (Rock *et al.* 1987, McNeil and Kerrich 1986, Marmont and Corfu 1988). Consistent with this premise is the apparent elevated concentrations of gold in some lamprophyre dikes (Rock *et al.* 1987) and the possible higher degree of oxidation of silica-under-saturated magmas in the Abitibi belt (Cameron and Carrigan 1987). The data required to characterize these factors in all intrusions, regardless of chemical affinity, proximity to gold mineralization, and age of emplacement, are not available; hence, suggestions that silica-undersaturated magmas were genetically related to gold remain speculative.

Insufficient data are available to exhaustively assess differences in magmatic fluid evolution, related to differences in magma composition, beyond the influence of H_2O and CO_2 present in the source region or the solubility of volatiles (eg. Burnham 1979). Based on fluid inclusion studies, CO_2 can be a common component in the upper mantle (eg. Touret and Dietvorst 1983). Magmatic fluids evolved from silicate melts (eg. silica-under-saturated), generated by partial melting in the presence of CO_2 , could become CO_2 -rich (see above).

2.3.2. REDOX CONDITIONS

Different oxidation states may exist in felsic magmas of different origin. I-type magmas are more oxidized than S-type (Ishihara 1975, 1981). Examples of extrusive, oxidized volcanic rocks are provided by the Taupo rhyolites (Ewart *et al.* 1971), whereas most basaltic rocks (Haggerty 1978) and the Yellowstone rhyolites (Christiansen and Blunk 1972) equilibrated near the quartz - fayalite - magnetite (QFM) buffer. While it was argued above that the role of source rock in Archean gold metallogenesis

PART II. GENETIC MODEL

is difficult to evaluate, it is possible that source rock had an influence on fO_2 of a derivative partial melt. The conspicuous presence of silica-undersaturated intrusions in the late plutonic suite, locally in close proximity to gold deposits which show evidence of having developed from relatively oxidized fluids (Cameron and Hattori 1987), leads to the inference that this magmatic suite may have been relatively oxidized (Cameron and Carrigan 1987). Conversely, intrusions of the TTG suite, which is the characteristic magma suite of most of the plutonic history of the Archean greenstone belts, were generated by 10 to 20 % partial melting of amphibolite (eg. Arth and Barker 1972, Arth *et al.* 1978, Barker and Arth 1976), and may have evolved under fO_2 conditions buffered by the QFM buffer, given the basaltic bulk composition of the protolith. Thus, by virtue of their differing source rocks, the fO_2 in the source region for TTG suite may have been lower (less oxidizing) than that where silica-undersaturated magmas were generated.

Exsolved vapour in equilibrium with each silicate melt type would assume the fO_2 of its parental melt, which is buffered mainly by Fe^{2+}/Fe^{3+} ratios. During ascent and cooling, magmas and exsolved magmatic fluids in equilibrium with crystals and melt would follow temperature - fO_2 paths parallel to major redox buffers (Mathez 1984). On separation from the melt, the fO_2 of exsolved fluids would be buffered internally, perhaps by the relative amounts of oxidized (SO_2) and reduced (H_2S) S-bearing compounds. Under relatively oxidizing fO_2 conditions, SO_2 would be the dominant S-bearing compound in the magmatic fluid and the fluid might cool along a temperature- fO_2 path parallel to the H_2S/SO_2 curve (Cameron and Hattori 1987, Ohmoto and Rye 1979), moving towards the hematite - magnetite buffer.

2.3.2.1. Implications for Gold and Sulphur Availability

The speculation that some late plutons and magmatic fluids may have originated under relatively oxidizing conditions has some important implications on the availability of gold and sulphur in relatively oxidized, hydrothermal fluids. The mechanism of solution for SO_2 in aluminosilicate melts is similar to that of CO_2 , although complicated by the at low values of fO_2 (Haughton *et al.* 1974). Sulphur is more soluble in oxidized silicate melts (Figure 2.3; Carroll and Rutherford 1985). This greater capacity to dissolve sulphur would be reflected in the sulphur abundance in an exsolved, magmatic fluid, a feature

confirmed by the S-rich nature of ore deposits associated with I-type magmas (Ishihara 1977, 1979). If Archean gold-related fluids evolved from oxidized magmatic hydrothermal systems, then they should also be relatively sulphur-rich. Virtually all Archean gold systems are characterized by the ubiquitous presence of sulphide minerals, notably pyrite. It is possible that gold solubility in an oxidized silicate melt or fluid was favoured by higher oxygen fugacity, because of its oxidation from Au^+ to Au^{3+} (eg. Fyfe and Henley 1973).

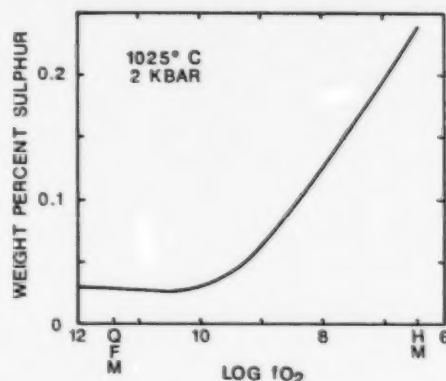


Figure 2.3. Solubility of sulphur in silicate melts, as a function of fO_2 . Data from Carroll and Rutherford (1985).

2.4. HYDROTHERMAL FLUID TEMPERATURE

The inferred trapping temperature for fluids related to gold precipitation (200° to 400°C; clustered about 300° to 350°C) is much lower than that for magmatic fluids related to porphyry Cu - (Mo) mineralization (400° to 700°C), and is 200° to 300°C lower than the granite solidus under conditions of excess H_2O (eg. Wyllie 1977). This significant temperature difference can be attributed to irreversible, adiabatic expansion of the fluid, which can be accompanied by an "instantaneous" drop in fluid temperature of as much as 300°C (Westra 1979, Macdonald 1983). Thus, the temperature difference between the gold-related fluids and the vapour excess solidus for a granitic magma is not inconsistent with a magmatic model of fluid generation.

2.5. JUVENILE FLUIDS

High density, CO_2 -rich fluids are commonly observed in upper mantle rocks (Touret 1971); therefore, the mantle provides an additional, large

source of CO₂-bearing fluids (Gold and Soter 1980, Newton *et al.* 1980). However, few data are available that provide a reliable characterization of juvenile fluids, particularly during the Archean. Therefore, it must be assumed that some CO₂-bearing magmatic fluids could have been provided from the mantle, either directly by degassing or by exsolution from mantle-derived silicate magmas (Frost and Frost 1987).

2.6. SUMMARY

Different magmatic fluid evolutionary paths are expected, dependent in part on initial magma composition, redox conditions at magma source, and the presence of H₂O and CO₂ at the magma source. Magmatic fluids, having a constitution similar to that of the gold-related fluids, could be produced by classical magmatic exsolution. A spectrum of fluid constitutions could result, dependent on the depth at which exsolution took place, the degree to which the exsolved fluid remained in equilibrium with the silicate magma, and whether the fluid underwent immiscible separation.

In the Abitibi belt, silica-undersaturated intrusions are spatially (eg. Todd 1928, Thomson *et al.* 1948, McNeil and Kerrich 1986) and temporally (Marmont and Corfu 1988) associated with gold. At least some of these silica-undersaturated intrusions were relatively oxidized, and appear to have

evolved an oxidized ore-forming fluid (Cameron and Carrigan 1987, Cameron and Hattori 1987). Conversely, the TTG suite may have originated under less oxidizing conditions by virtue of having been generated by partial melting of amphibolitic source rocks. Thus, this magma suite may not have been generated under suitably oxidizing conditions to develop a S-rich, oxidized silicate magma or magmatic fluids. However, no thorough characterization of magma oxidation state is presently available to document the presence and magnitude of redox variations within and between Archean magma suites. This remains an extremely interesting working hypothesis (Cameron and Carrigan 1987, Cameron and Hattori 1987), which will be addressed systematically and exhaustively in Ontario.

Different states of magma oxidation may have played a significant role in determining the ability of a magmatic fluid to extract and carry gold. Preliminary data suggest that certain Archean silicate magmas and auriferous hydrothermal fluids originated under relatively oxidizing conditions, such that SO₂ was the dominant S-bearing fluid species (Cameron and Hattori 1987). The subsequent cooling history of such a fluid, dependent primarily on the extent to which the exsolved fluid remained coupled and in equilibrium with silicate melt - crystal parent, determined if SO₂ or H₂S dominated when the fluid arrived at the depositional site.

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3. METAMORPHIC FLUIDS

3.1. INTRODUCTION

Derivation of hydrothermal fluid by prograde devolatilization reactions has been proposed as a process responsible for the generation of the Archean auriferous fluid (Kerrick and Fyfe 1981, Groves and Phillips 1987). In this chapter, some general characteristics of metamorphic fluids are reviewed to assess which, if any, metamorphic processes could generate a fluid akin to that implicated in the gold-forming event. This chapter is, of necessity, short, due to inadequate data on the topic. This is not intended as an exhaustive review of the reactions which generate fluid during prograde metamorphism, nor are the conditions under which metamorphic fluids evolve addressed. Many such reviews are available elsewhere (eg. Fyfe *et al.* 1978). While an exhaustive characterization of the composition of metamorphic fluids as a function of metamorphic grade, is not as well documented as ore-forming fluids, some generalizations are summarized by Crawford (1981) and Touret and Dietvorst (1983).

Three characteristics of metamorphic fluids relevant to Archean gold metallogeny are:

- (1) Fluids generated by dehydration reactions up to and including amphibolite metamorphic grade are compositionally variable, but tend to be H₂O-rich.
- (2) Ambient fluids present in the zone of granulitization and/or anatexis are CO₂-rich, regardless of host rock composition.
- (3) Fluids generated within a lithological domain, at a particular metamorphic grade, are in equilibrium with that rock assemblage, but may not be in equilibrium with a different rock assemblage under the same metamorphic conditions, or the same rock assemblage at lower metamorphic conditions.

The relevance of these characteristics to gold metallogeny is discussed in the following sections.

3.2. LOW TO MEDIUM METAMORPHIC GRADES

In low to medium grade metamorphic terranes, metamorphic fluids are dominated by H₂O, generated by dehydration reactions and/or mobiliza-

tion of pore fluids and intergranular fluid films (Fyfe *et al.* 1978). Fluids are generated in pulses, the volume of which are related partly to the thermal gradient and the bulk composition of the rock (Fyfe *et al.* 1978). Variable amounts of CH₄ and CO₂ may be generated during prograde metamorphism, dependent to some extent on the composition of the immediate country rock. Above the staurolite isograd, ambient metamorphic fluids contain some CO₂, with abundances of 10 to 60 mole percent.

The most significant characteristic of metamorphic fluids in the upper crust is their compositional dependency on the bulk composition of the source rock which is undergoing devolatilization reactions (eg. CH₄-rich fluids in carbonaceous sediments or NaCl brines related to evaporites; Rich 1979). This suggests that fluid composition is strongly dependent upon and buffered by the composition of the country rock (internal buffering; Touret and Dietvorst 1983).

3.3. HIGH METAMORPHIC GRADES

The transition into high grade metamorphic rocks, where partial melting may occur, corresponds to a major change in ambient metamorphic fluid constitution, from H₂O-dominant under lower metamorphic facies to CO₂-dominant in high grade terranes (Figure 3.1; Touret 1981, Touret and Dietvorst 1983). In granulites, the ambient fluid during peak metamorphism consists of high density (up to 1.23 g/cm³) CO₂, regardless of local host rock composition. Other CO₂ -, H₂O - CH₄ or N₂ - bearing fluid inclusion populations are present and can often be attributed to post-peak metamorphic retrograde events (Touret 1981, Touret and Dietvorst 1983).

The presence of CO₂ in rocks of variable composition in granulite facies terranes points to an external control of the fluid composition (external buffering; Touret 1981, Touret and Dietvorst 1983). At least two mechanisms have been proposed to account for the "external buffering": (1) CO₂-rich fluid was introduced into the lower crust and decreased the activity of H₂O (eg. Newton *et al.* 1980, Wells 1979), and (2) CO₂ was indigenous to the lower crust, but its concentration increased because H₂O was preferentially partitioned into and removed by silicate partial melts (eg. Touret and Dietvorst 1983). Evidence for both mechanisms will be discussed in chapter 4 of Part II.

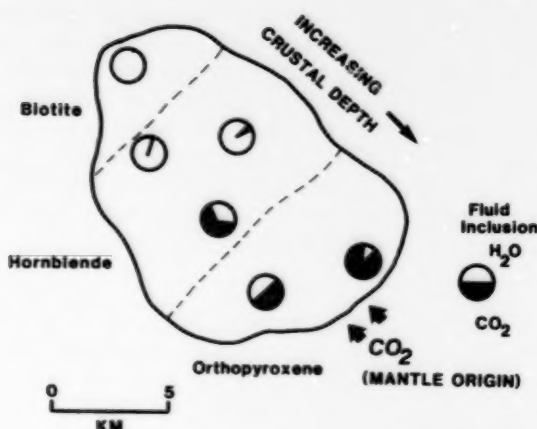


Figure 3.1. Observed change in metamorphic fluid composition, across the transition from medium to high metamorphic grade. CO_2 -rich fluids dominate in the high grade metamorphic terranes, regardless of rock composition. Data from Touret (1981).

3.4. SALINITY

Salinity of metamorphic fluids is variable, but seems to reflect the composition of the local rock. At low to medium grade, in terranes where carbonate or evaporitic rocks are absent, salinity of ambient metamorphic fluids is generally less than 2.4 wt % NaCl equivalent, and variation can not be correlated with metamorphic grade (Crawford 1981). In higher grade metamorphic terranes, the salinity of metamorphic fluids is generally low (Crawford 1981).

3.5. RELEVANCE TO ARCHEAN GOLD-RELATED FLUIDS

Four salient fluid characteristics of the Archean, gold-related hydrothermal systems are:

- (1) The fluid was aqueous, of relatively uniform composition, and contained CO_2 (Chapter 6, Part I).
- (2) The fluid was derived from a source external to the immediate depositional site (Chapters 1, 2 and 3, Part I).
- (3) A large volume of fluid was involved (Chapter 3, Part I).
- (4) Extensive alteration at the depositional site indicates that the fluid was not in equilibrium with the rock assemblage at the depositional site.

librium with the rock assemblage at the depositional site. By implication, the gold-related fluid was not in equilibrium with the rock assemblage in the metamorphic domain where the gold was precipitated (Chapter 3, Part I).

3.5.1. LARGE FLUID RESERVOIR OF UNIFORM COMPOSITION

Ambient metamorphic fluids during high grade metamorphism (granulitization) are compositionally uniform, regardless of rock composition. Similarly, the composition of gold-related hydrothermal fluids was also relatively uniform, regardless of deposit location or bulk composition of the host rock. This may indicate that the control on the composition of the gold-related hydrothermal fluid may lie in the granulitization of the lower crust. Further controls on fluid composition and its association with granulitization are shown by the temporal similarity between the gold precipitation event and lower crustal granulitization. This is demonstrated in Chapter 4, Part II. This is not evidence that the gold-related fluids were identical in composition to those ambient in the zone of granulitization, which are CO_2 -rich compared to the gold-related fluids. However, the zone of granulitization provides a very large, compositionally uniform fluid reservoir, from which the gold-related fluid may have evolved.

3.5.2. DISEQUILIBRIUM

Metamorphic fluids generated by dehydration reactions in low to medium grade metamorphic terranes are in equilibrium with and have a composition buffered by mineral assemblages in the rock. The stable assemblages in metamorphic rocks are universally typical. The localized transformation of regionally extensive hydrous assemblages, typical of low to medium grade metamorphic terranes, to the localized carbonate- and aluminosilicate-rich assemblages typical of Archean gold systems (Chapter 3, Part I) is evidence that the gold-related fluids were not in equilibrium with the country rocks when first introduced to the depositional site. This implies that the gold-related fluid was not generated within the metamorphic domain which hosts the gold mineralization. Given that Archean gold mineralization is developed in rocks of greenschist and amphibolite metamorphic grade, this suggests that the fluid was generated under conditions different from those of greenschist and amphibolite metamorphic facies. High grade metamorphic terranes satisfy this requirement. This indicates that, if the gold-related

hydrothermal fluid was of metamorphic origin, it originated from a higher grade metamorphic terrane - namely, the zone of granulitization.

3.5.3. CO₂-BEARING FLUID

The persistent CO₂ abundance of the gold-related fluid requires a source which contains a uniformly distributed supply of carbonate or CO₂. Dispersed calcite, precipitated during sea water alteration of the volcano-sedimentary assemblage (Chapter 6, Part I), may provide some CO₂ to a metamorphic fluid. However, the carbon isotopic data are not consistent with this model (Chapter 1, Part II). Dissolution of hydrothermal carbonate from regionally distributed and fault-related zones of carbonatization could also provide CO₂ to a metamorphic fluid, having a $\delta^{13}\text{C}$ composition similar to that of the gold-related fluids. However, the introduction of this carbonate was coeval with gold introduction. Therefore, the sole metamorphic reservoir which contained a large volume of CO₂ and was external to the metamorphic terranes where gold is localized was the zone of granulitization.

3.6. METAMORPHIC OR MAGMATIC FLUID

On the basis of the compositional uniformity of the gold-related hydrothermal fluid, in spite of the variety of host rock compositions, the fluid is interpreted to have been generated outside the domain which contains the gold deposits. Compositional uniformity and ubiquitous CO₂ content suggest that the gold-related fluid was generated by a regionally uniform mechanism which produced abundant CO₂-rich fluid. A granulite facies source is considered the most probable metamorphic reservoir from which an auriferous fluid may have evolved.

Geochronological support for this working hypothesis is presented in the following chapter. However, as alluded to above, the CO₂-rich nature of the ambient fluid in the zone of granulitization may reflect magmatic processes, either by introduction of juvenile CO₂ or enrichment of CO₂ as a result of extraction of H₂O during partial melting. An important conceptual problem is posed: is the fluid in the zone of granulitization metamorphic or magmatic? This question will be addressed in Chapters 4 and 5 of Part II. However, regardless of the metamorphic processes, we conclude that granulitization may have been involved in the Archean gold metallogenic event.

4. ARCHEAN CRATONIZATION AND GOLD INTRODUCTION

4.1. INTRODUCTION

Across the Superior Province, a systematic and similar, magmatic and tectonic cratonization sequence is recorded in all subprovinces. However, this sequence of events occurred at different times in the various subprovinces (Stott *et al.* 1987). Two well constrained igneous events in the Superior Province are the dates of cessation of calc-alkaline volcanism and related sedimentation, and the following, relatively short period, when late, post-volcanic, silica-saturated and -undersaturated intrusions were emplaced (Figure 4.1, Part II). Anomalous concentrations of gold and large volumes of CO₂ were introduced into regionally extensive zones of deformation within the upper crust during that period of late plutonism (Chapter 4, Part I).

Examined in this chapter is the late Archean magmatic and thermal record of the Superior Province. This is achieved by comparing the late magmatic evolution of the upper crust in the southern Superior Province (Abitibi - Wawa and Wabigoon Subprovinces) with the late thermal history of the lower crust, as represented by the high grade (granulite facies) rocks exposed in the Kapuskasing Structural Zone and parts of the Winnipeg River Subprovince. An important result of this analysis is the recognition that the introduction of gold was but one of a sequence of coeval magmatic, tectonic and thermal events which collectively affected the Archean crust of the Superior Province.

An integrated cratonization and gold metallogenic model is developed using empirical, geochronological, and geochemical characteristics of Archean terranes in the Superior Province, to account for the remarkable contemporaneity between magmatic, metamorphic and auriferous hydrothermal events.

4.2. MAGMATIC HISTORY OF THE UPPER CRUST - SOUTHERN SUPERIOR PROVINCE

U-Pb zircon ages for volcanic and intrusive rocks from the Kirkland Lake, Timmins, Wawa, Gamitagama, and Hemlo areas and from the Wabigoon Subprovince are summarized in Figure 4.1. In the southern part of the Abitibi - Wawa Subprovince, felsic volcanic rock and contemporaneous plutonism had been initiated by 2750, and possibly

as early as 2772 Ma., but had terminated by 2700 Ma. The emplacement of compositionally diverse plutons (between 2699 and 2666 Ma.; Chapter 4, Part I) continued following the termination of calc-alkaline volcanism (Nunes and Pyke 1980, Nunes and Jensen 1980, Krogh and Turek 1982, Turek *et al.* 1982, Percival and Krogh 1983, Corfu and Stott 1986, Smith *et al.* 1987, Marmont and Corfu 1988, Corfu and Muir 1988). In the Wabigoon Subprovince, volcanism, related sedimentation and contemporaneous plutonism had been initiated by 2775 Ma. and terminated at approximately 2700 Ma. Post-volcanic plutonism continued locally until 2675 Ma. (Davis and Edwards 1982, 1986, Davis *et al.* 1982, Davis *et al.* 1988, Smith *et al.* 1988).

4.2.1. GOLD AND CO₂ INTRODUCTION

The introduction of large volumes of CO₂ into the greenstone belts is an inherent characteristic of the late Archean gold metallogenic and deformation events. As discussed in Chapter 6 of Part I, a large volume of dolomite - ankerite carbonate was introduced into late, regionally extensive structures, coevally with the gold event. Thus, implications regarding the absolute time of gold introduction (see Chapters 4 and 5, Part I) also apply to CO₂ introduction.

A critical aspect of the gold metallogenic event in the southern Superior Province is the date of gold concentration into regional zones of deformation. In the Timmins area, gold was introduced after 2673 Ma. (Marmont and Corfu 1988). A lower limit on hydrothermal activity within auriferous systems is provided by ⁴⁰Ar/³⁹Ar dates for hydrothermal micas that occur within and immediately adjacent to gold-bearing veins in the Dome and Hollinger Mines (2633 Ma. and 2617 Ma. respectively, Masliwec *et al.* 1986). In the Hemlo camp, the age of gold mineralization and related hydrothermal activity is constrained to the period 2690 to 2670 Ma. (Corfu and Muir 1988), although evidence for younger hydrothermal activity in the auriferous zone is provided by ⁴⁰Ar/³⁹Ar dates for white micas (maximum age of 2671 Ma.; Masliwec *et al.* 1986) and U-Pb ages of monazite and rutile (2650 - 2610 Ma.; Corfu and Muir 1988).

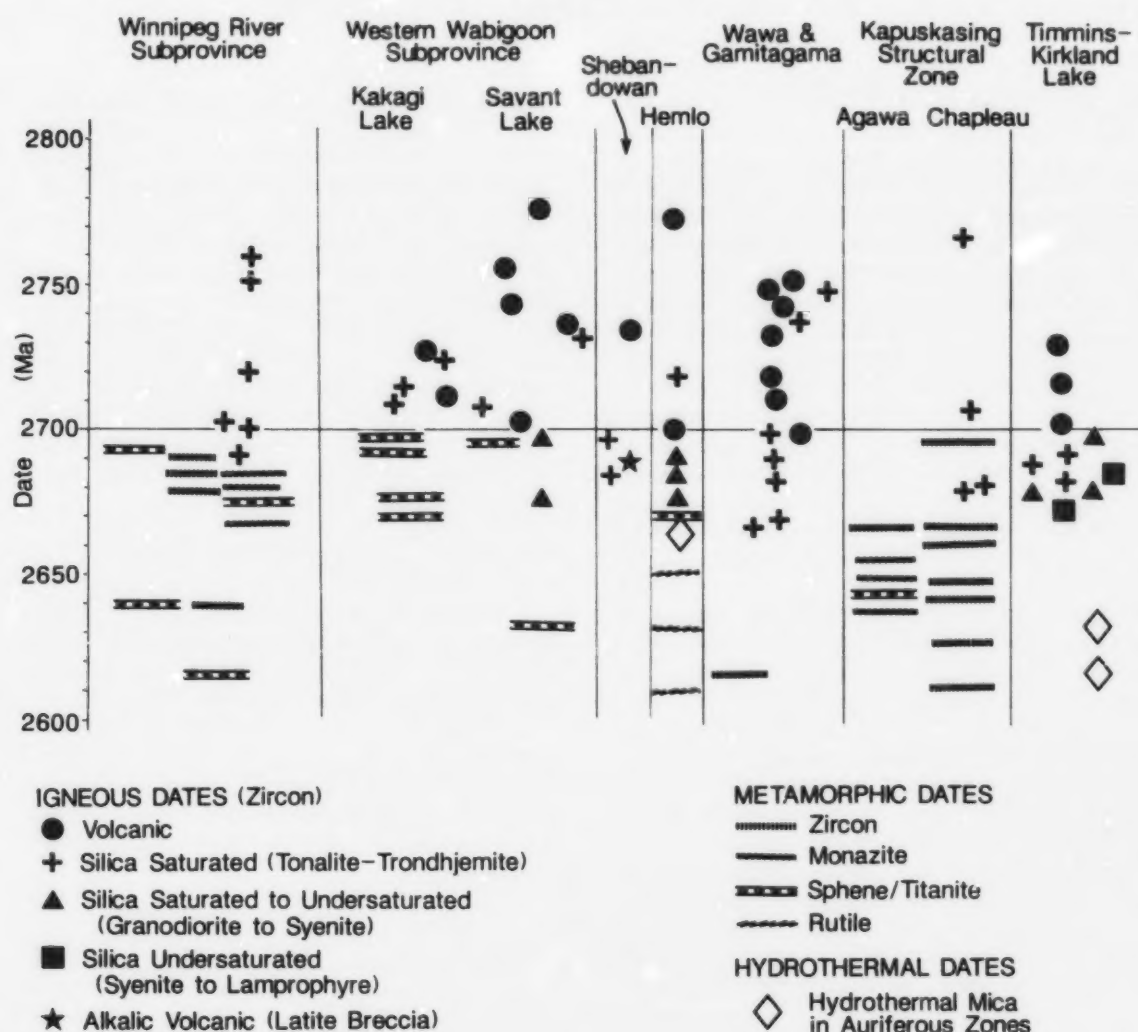


Figure 4.1. Summary of geochronological data for parts of the southern Superior Province, Ontario. Sources of data: Timmins and Kirkland Lake - Nunes and Pyke (1980), Nunes and Jensen (1980), Marmont and Corfu (1988); Kapuskasing Structural Zone and Wawa - Percival and Krogh (1983), Turek et al. (1982), Krogh and Turek (1982), Percival and Krogh (1983), Krogh et al. (1986), Smith et al. (1987), Corfu (1987), Krogh et al. (1988), Percival et al. (1988); Shebandowan Belt - Corfu and Stott (1986); Hemlo Belt - Corfu and Muir (1988); Wabigoon Subprovince - Davis and Edwards (1982, 1986), Davis et al. (1982), Davis et al. (1988); Winnipeg River Subprovince - Corfu (1988, in press).

4.3. THERMAL HISTORY OF THE LOWER CRUST

A cross section through the upper two-thirds of Archean crust beneath the Superior Province is provided by the Kapuskasing Structural Zone (Percival and Card 1985), and by parts of the Winnipeg River Subprovince which have attained granulite facies metamorphic grade. The temporal development of lower crustal metamorphism is pro-

vided by U-Pb geochronology of various mineral phases, including zircon, monazite, titanite and rutile. It is emphasized that ages of these minerals represent the time when the minerals cooled through their "blocking temperature". Thus, equilibrium mineral assemblages may have been attained at an older age than closure.

In the Agawa migmatite terrane, at the south end of the Kapuskasing Structural Zone

(KSZ), metamorphism with attendant magmatism and migmatization culminated between 2670 and 2665 Ma. with an episode of partial melting (Corfu 1987, Figure 4.1). This was followed by injection of magma, which caused an increase in temperature and probably induced further ductile deformation and metasomatic reactions with country rocks at about 2663 - 2660 Ma. (Corfu 1987). These thermal events concluded with the intrusion of granites and pegmatites and the initiation of extensive, late hydrothermal activity within the lower crust between 2660 and 2635 Ma. (Corfu 1987).

Ductile deformation and granulite to amphibolite facies metamorphism in the main part of the KSZ (Chapleau terrane) was attained between 2696 and 2650 Ma. (Percival and Krogh 1983, Krogh *et al.* 1986, Krogh *et al.* 1988, Percival *et al.* 1988). The dates obtained from granulite facies rocks decrease systematically to deeper levels of the crust; this is interpreted to represent slower cooling at deeper crustal levels (Percival and Krogh 1983, Krogh *et al.* 1986, Percival *et al.* 1988).

In the Winnipeg River Subprovince (WRS), granulite facies metamorphism, accompanied by local magmatism, peaked at about 2680 Ma. (Corfu 1988). Areas affected by this high grade metamorphic event were subjected to a prolonged period (<2680 Ma.) of lower grade, retrograde metamorphism and metasomatism (Corfu 1988).

4.3.1. HYDROTHERMAL ACTIVITY IN THE LOWER CRUST

Locally within the KSZ and WRS, hydrothermal fluids circulated and disrupted the U-Pb systems in various minerals for at least 60 Ma. after the granulite facies metamorphism (Corfu 1987, Corfu 1988, in press, Krogh *et al.* 1988). Evidence of this activity includes the presence of morphologically distinct, high-U zircon overgrowths on pre-existing, low-U zircon cores (KSZ - Krogh *et al.* 1988, Corfu 1987), conflicting age and paragenetic relationships (KSZ - Corfu 1987; WRS - Corfu 1988), and irregular titanite and apatite ages within small areas affected by the granulite facies metamorphism (Corfu 1988). These data are interpreted to reflect the influence of hydrothermal fluids, and are not consistent with simple cooling histories (Corfu 1988, 1987). The late "hydrothermal" zircons have morphological and chemical characteristics consistent with their lower temperature formation in fluid-rich environments, containing incompatible elements such as Zr, U and Th (Corfu 1987). Uran-

iferous pegmatites and granodiorite to granitic plutons, emplaced between 2660 and 2635 Ma. in the Agawa migmatite terrane (KSZ), and pegmatitic intrusions in the southern part of the WRS, may have been generated during this late hydrothermal episode (Corfu 1987).

4.4. LATE MAGMATIC AND THERMAL HISTORIES OF UPPER AND LOWER CRUST - A COMPARISON

Compositionally diverse, post-volcanic plutons were emplaced into the supracrustal rocks of the southern Superior Province between 2700 and 2666 Ma., coeval with the high grade metamorphism and attendant partial melting of the lower crust in the KSZ and WRS (2696 - 2630 Ma.; Figure 4.1). Gold introduction and related hydrothermal activity in the Timmins and Hemlo camps (<2673 Ma. and 2690 - 2670 respectively) also took place towards the end of or following the late plutonic event, contemporaneous with the lower crustal, granulite facies metamorphism. Thermally anomalous conditions existed in the Timmins and Hemlo auriferous systems until 2610 Ma. (white mica, rutile and monazite dates; Figure 4.1), coeval with the post-peak metamorphic thermal history of the lower crust. These geochronological data illustrate that the late, magmatic and thermal histories of the upper and lower crust in the southern Superior Province were temporally equivalent and that the introduction of anomalous concentrations of gold and CO₂ into the upper crust was an integral component of this coupled, late magmatic and thermal event.

4.5. CATALYST TO LATE CRUSTAL HISTORY

The coeval generation of the late plutons and the granulitization of the lower crust may indicate that both were initiated by, and are manifestations of, a similar event. Many models of granulitization appeal to an external source of heat or fluid to induce granulitization (eg. Bohlen 1987, Touret and Dietvorst 1983, Frost and Frost 1987). Thus, it is reasonable to speculate that the late granulite metamorphism of the lower crust and the emplacement of magmas of diverse composition were triggered by a process external to the crust. Furthermore, it has been suggested that granulitization of the lower crust may have played an important role in Archean gold metallogeny (Fyon *et al.* 1983b, 1984, Cameron 1988). Consideration of the mechanisms by which granulitization is effected may provide clues to the existence and nature of the pro-

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cess capable of triggering the late magmatic and thermal history of the crust.

4.6. GENERAL CHARACTERISTICS OF GRANULITES AND MODEL OF GRANULITIZATION

While an exhaustive review of the process is beyond the scope of this report, granulitization appears to require the presence of and result in the generation of CO₂-rich metamorphic fluids (eg. Touret and Dietvorst 1983; Chapter 3, Part II). Two conventional interpretations of granulitization include: 1) granulites represent a residuum from which a felsic melt has been removed (eg. Fyfe 1973, Pride and Muecke 1980), and 2) granulites were produced by reaction between amphibolitic rock and CO₂-bearing volatiles, which may or may not involve localized anatexis (eg. Hamilton *et al.* 1979, Newton *et al.* 1980). Many low and medium pressure granulite rocks are depleted in certain LIL elements, such as K, Rb, Cs, U, Th, Y, heavy rare earth elements, and have high Ba/Rb, Ba/Sr, Ce/Yb, K/Cs and K/Rb ratios, relative to inferred protoliths (eg. Rollinson and Windley 1980b). For these chemical characteristics to be attributed to partial melting phenomena, the derivative melt must be geochemically complementary, and would be characterized by low K/Rb, Ba/Rb, and Ba/Sr ratios, and would be relatively enriched in Rb, because Rb is preferentially depleted from granulites with respect to K and Ba (Drury 1973).

Granulite development has also been attributed to equilibration between amphibolitic protoliths and mantle-derived (external), CO₂-bearing fluids (eg. Schuiling and Kreulen 1979, Newton *et al.* 1980, Frost and Frost 1987). This is based on the presence of CO₂-rich fluid inclusions in metamorphic minerals (Berglund and Touret 1976, Touret 1971a, b, 1977, Hoefs and Touret 1975, Touret and Dietvorst 1983) and the patchy, vein-like appearance of charnockite, observed in the transition zone from amphibolite to granulite facies rocks (Janardhan *et al.* 1982).

Thermal characteristics of high metamorphic grade gneiss terranes provide some additional constraints on the granulitization mechanism. The thermal history of some granulites requires an anti-clockwise pressure - temperature - time (P-T-t) path evolution (Bohlen 1987), during which peak temperature is attained before maximum pressure (Bohlen 1987). This thermal condition can result from the introduction (underplating) of magmas at the base

of the crust, the intrusion of magmas into the crust, and/or the passage of magmas through the crust (Wells 1980), but is not consistent with thermal models of continent - continent collision (England and Thompson 1984).

Thus, the chemical composition of some low to medium pressure granulites and their deduced thermal history (anti-clockwise P-T-t path) suggest that some granulite terranes may have developed in response to the introduction of magma to the base of and into existing continental crust, perhaps by the process of magma underplating (Bohlen 1987, Frost and Frost 1987). The attendant localization of heat and volatiles (H₂O and CO₂) catalyzed the heating, dehydration and partial melting of the lower and middle crust (Newton *et al.* 1980, Touret and Dietvorst 1983, Frost and Frost 1987).

This is only one possible model of granulitization. Granulites can be generated during the process of continent - continent collision (Wilks 1988), and therefore the generation of granulites does not necessarily require the addition of anomalous heat from externally derived magmas. However, the evidence provided from the interpreted pressure - temperature - time paths of some granulites (Bohlen 1987) is more consistent with the magma underplating model. The underplated magma could be generated either by subduction-initiated partial melting or other mantle initiated diapirism. This aspect of the model is addressed in the following chapter.

4.7. CRATONIZATION OF THE SOUTHERN SUPERIOR PROVINCE

In this section, the cratonization of the Superior Province is tested by determining the coherency between the late magmatic and thermal histories of the southern Superior Province and those predicted by the underplating model.

4.7.1. MANTLE DIFFERENTIATION

A significant differentiation of the primitive Earth's mantle is inferred to have culminated between 2800 Ma. and 2500 Ma., during which some mantle domains became isotopically isolated (Bell *et al.* 1982). The differentiation is interpreted to reflect the ascent and partial melting of mantle diapirs, which originated in the asthenosphere and/or mesosphere (Brooks *et al.* 1976). Such "primitive" mantle material is inferred to exist

beneath part or all of the Superior Province, and has acted as a source for volumetrically minor amounts of radiogenically distinct, Phanerozoic, carbonatite magma (Bell *et al.* 1982). Direct mantle contribution of magma to the upper crust during the late Archean is postulated on the basis of major and trace element composition, and radiogenic isotope (Sm/Nd) characteristics of some volcanic and plutonic rocks in the western part of the Superior Province (Shirey and Hanson 1984). Some of these igneous rocks have compositions similar to sanukites, interpreted to represent direct partial melting of the mantle (Shirey and Hanson 1987, Tatsumi and Ishizaka 1982, Tatsumi 1982). Thus, mantle domains beneath parts the Superior Province may have undergone a major partial melting and devolatilization event during the Late Archean, which culminated with the coupling of mantle and the crust at about 2800 - 2500 Ma.

The inferred coupling between mantle and crust, and the supply of mantle-derived magma into (and underplating ?) the crust, may have been the catalyst which granulitized the lower crust. Numerical modeling of the cooling history for the KSZ indicates extremely low cooling rates for the granulites, from 800° to 1000°C at the metamorphic peak, with initially high heat flow (Percival *et al.* 1988), consistent with the underplating model.

4.7.2. LATE CRUSTAL PLUTONISM

Late emplacement of compositionally diverse plutons into the upper crust (2700 - 2675 Ma.) was coeval with lower crustal granulitization (2690 - 2630 Ma.), and is consistent with the model prediction that granulite facies metamorphism should be accompanied by a pulse of plutonism (Bohlen 1987). Furthermore, magmas produced during this late plutonic event were derived from both crustal and mantle sources (eg. Shirey and Hanson 1984). Some silicate intrusions may have been derived by partial melting of lower crust, in the zone of granulitization. Pegmatites, which intrude along the southern boundary of the Wabigoon Subprovince, have low K/Rb and Ba/Rb ratios (Figure 4.2), which complement the depleted geochemistry of granulitized terranes. Although not characterized geochemically, uraniferous pegmatites within the Agawa migmatite terrane may also have formed by a similar process (Corfu 1987). However, plutonic complexes with varying degrees of silica-saturation, displaying such chemistry, have not been recognized in the Superior Province (Figure 4.2). Some intrusions belonging to the TTG suite could have been

derived from either garnet-bearing amphibolite (either underplated, mafic lower crust or greenstone belt roots) or garnet-lherzolite (eg. Martin 1987, Rollinson and Windley 1980a).

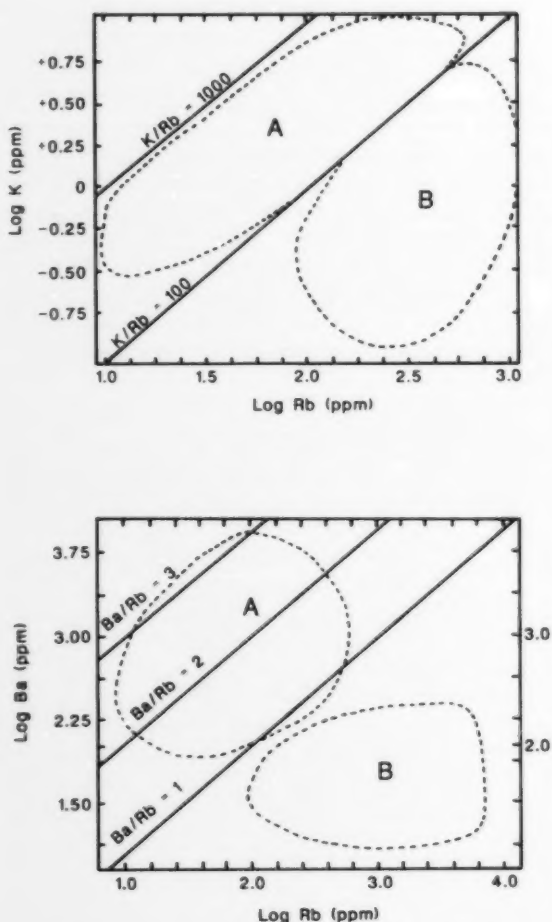


Figure 4.2. Log-log plot of K/Rb and Ba/Rb for intrusive rocks from the parts of the Uchi, Wabigoon and Abitibi Subprovinces. Field A: silica-saturated and -undersaturated intrusions. Field B: silica-saturated and -undersaturated intrusions and pegmatitic rocks from the Wabigoon Subprovince which may have been generated by partial melting in the zone of granulitization. Data from Ontario Geological Survey, PETROCK data base.

4.8. CRATONIZATION MODEL

One possible model for the cratonization of the Shield which accounts for the well constrained magmatic and thermal history of the southern part

solidus and higher viscosity relative to its volatile-bearing precursor (Pollack 1986). If the partial melting and devolatilization accompanied the coupling event, a subcontinental "keel" may have grown by the addition and "freezing" of diapiric magma to the base of the crust (underplating). This material would be more refractory, have a higher viscosity and be seismically anomalous with respect to mantle and crustal domains which had not experienced a similar history. Thus, the seismically anomalous mantle zone beneath the Red Lake area may be a consequence of the underplating event.

4.10. SUMMARY

The consistency between the late magmatic and thermal events of the upper and lower crust (2700 - 2650 Ma.) is a manifestation of the cratonization of the Superior Province. The introduction of mantle-derived magma into the crust at this time suggests that underplating of mantle-initiated diapirs may have an important role in the cratonization event. That anomalous concentrations of gold and CO₂ were introduced into the supracrustal packages during or the later stages of the late magmatic and thermal history illustrates that Archean gold metallogeny was inextricably linked to the late, magmatic and thermal history of the upper mantle and crustal systems.

Some terranes within the Superior Province (eg. parts of the Uchi and the western Wabigoon Subprovince) appear to have been cratonized before the southern Superior Province. This may indicate that the Superior Province shield consists of terranes or cratonic islands which experienced similar magmatic and thermal evolutionary sequences in response to mantle diapirism, but at different absolute times. Accretion of these terranes would form a thermally composite shield (Figure 4.4). Thus, rather than a single event during which gold was introduced into the Superior Province, at least two gold introduction events may have occurred, each triggered by the culmination of mantle-derived magmatism and its affect on sialic micro-cratons.

The appeal and strength of the mantle-derived magma underplating hypothesis is its holistic accounting of the late magmatic and thermal history of the Superior Province. As a "byproduct" of this cratonization process, anomalous concentrations of gold, CO₂ and other incompatible elements were introduced into the supracrustal sequences. Certain key empirical characteristics of the auriferous systems are also consistent with the underplating model. These are considered in the following chapter.

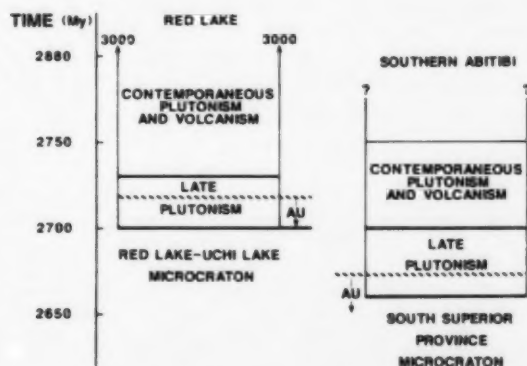


Figure 4.4. Cartoon illustrating the cratonization of "micro-cratons", through an equivalent sequence of igneous and thermal events. "Micro-cratons" are then accreted to produce the Superior Province. At least two events of gold introduction are recorded, although more may occur because the introduction of gold is a consequence of the cratonization process.

5. HOLISTIC MODEL OF ARCHEAN GOLD METALLOGENY

5.1. INTRODUCTION

A genetic model is distinguished from a depositional model in that it requires a significant degree of interpretation to develop a consistent explanation for a set of empirical observations. Late Archean crustal development records a history of temporally consistent upper and lower crustal magmatic, tectonic and thermal events, and argues for a cratonization process, during which gold was concentrated into specific structural sites within the upper crust. The basis of this genetic model, which was formulated in Chapter 4 of Part II, is that late magmatic and thermal aspects of the crustal history are attributed to the introduction of mafic magma, possibly of mantle origin, to the base of and into the crust. Many critical relationships required to constrain a genetic model remain ambiguous, particularly the role of mantle and crustal processes in influencing the style of late Archean tectonism.

5.2. GENETIC MODEL

Figure 5.1 presents a representation of the magmatic, tectonic and thermal state through the upper mantle and crust and is similar to Figure 4.3. The approximate crustal region where Archean gold mineralization is inferred to precipitate is illustrated. Figure 7.1 of Part I displays characteristics of this area in detail.

Hydrothermal CO_2 - H_2O fluids are generated by exsolution from silicate magmas in the mantle and lower and upper crust. CO_2 -rich vapour may accumulate in the lower crust, by virtue of preferred partitioning of the H_2O component into and removal by a silicate magma. This fluid may evolve compositionally, in response to changing redox conditions in the zone of granulitization, and may pass directly into the upper crustal rocks along permeable structural systems. Fluid unmixing in the lower crust would add an additional degree of variability to the fluid composition (Chapter 2, Part II). Gold is introduced into zones of deformation from fluids which accessed a large volume of rock in transit to the depositional site. The mode of fluid and gold transport is not definite, although a magmatic origin, in the broadest sense, is consistent with the available data. Distinction between magmatic and metamorphic fluids, particularly as regards their relative participation in the generation of the

auriferous fluid, is moot, especially at the crustal scale being considered.

5.3. UNRESOLVED PROBLEMS

Given the interpretative and speculative nature of this chapter, it is perhaps presumptuous to imply that any aspects of the model are known with certainty. However, by considering some key aspects of the model, an appreciation for the level of uncertainty associated with each component can be assessed.

5.3.1. MANTLE - CRUST LINK

Coeval with the crustal events, the mantle may have been undergoing a period of diapirism, with attendant thermal convection, devolatilization and partial melting. Neither the subduction nor convection mechanism of magma generation can be unequivocally discriminated at present. However, all that is required to maintain consistency within the model is that magma, which originated from below the continental crust, be supplied to the base of and into existing continental crust. Direct mantle contribution of magmas to the upper crust is indicated (Shirey and Hanson 1984) and underplating of mantle-derived diapirs may have produced a sub-continental keel. What triggered the mantle diapirism remains to be demonstrated, and whether subduction or a solely mantle-related process was instrumental in delivering magma to the base of the continental crust, must be assessed.

5.3.2. COMPOSITION OF PARTIAL MELTS

The observed compositional spectrum in late plutonic group, from silica-saturated to -undersaturated, may reflect an evolving $\text{CO}_2/\text{H}_2\text{O}$ ratio of a vapour phase in the region of partial melting (eg. Mysen and Boettcher 1975). The composition of magmatic fluid is dependent on the bulk composition of the parent silicate magma, and this may have a bearing on the ability of some fluids to extract and transport gold more efficiently than others. The composition of vapour ambient during partial melting and magma crystallization remains to be documented by examination of primary fluid inclusions, preserved in the late plutonic bodies, and in plutonic mineral chemistry variations.

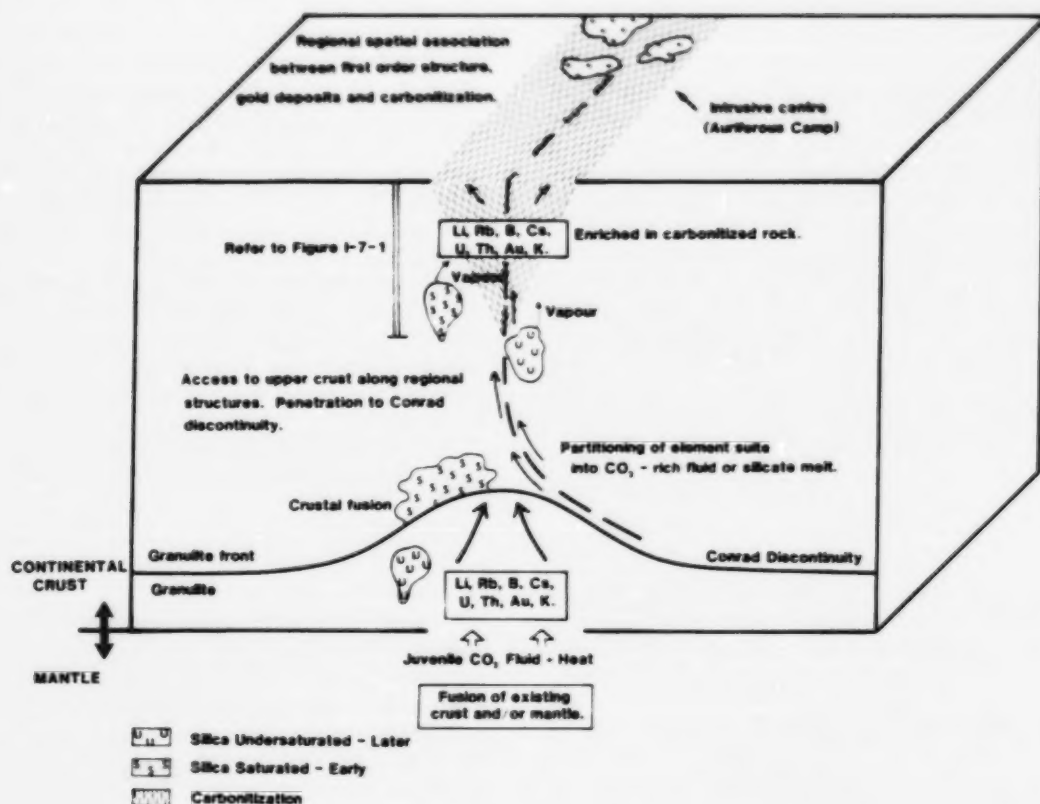


Figure 5.1. Composite crustal section illustrating the elements of an Archean gold genetic model. The indicated surface represents an erosion level, as observed presently. It is not intended to illustrate that the Archean auriferous hydrothermal system debouched onto the surface (see discussion in Chapter 7, Part I). The vertical scale is not defined. Crustal cratonization culminates with the addition of mafic magma to the base of and into existing lower crust. Crustal thickening is accomplished in part by the addition of diapiric mass to the base of the crust, which grows downward to form a sub-continental keel. Granulitization of the lower crust results from the introduction of magma, heat and volatiles, principally CO_2 . A compositional spectrum of silicate magmas, from silica-saturated to -undersaturated, is generated by partial melting of upper mantle and lower crust, in the presence of fluids having varying $\text{CO}_2/\text{H}_2\text{O}$ ratios. Fluids may originate by exsolution from silicate magmas anywhere between the upper mantle and upper crust, and may evolve compositionally as a result of reaction with rock (granulitization) and immiscible separation. Chemical mass transfer from the mantle, lower and middle crust to the upper crust is effected by silicate magma and fluids, which migrate along zones of crustal weakness (regional structures which penetrate at least to the Conrad discontinuity). Gold concentration in the upper crust is a consequence of this late crustal magmatic, tectonic and thermal history.

5.3.3. ROLE OF MAJOR CRUSTAL STRUCTURES

The role of the regional structures in this model is not completely understood. Mounting field (Hubert *et al.* 1984), geochemical, and geochronological (Davis *et al.* 1988) evidence supports structural stacking of volcanic and sedimentary packages during the tectonic history of the Canadian Shield. However, the regional structures may also represent boundaries between Subprovinces, across which sig-

nificant changes in pressure and temperature conditions of adjacent blocks are recorded. Integrated P - T studies, seismic reflection profiling, and detailed, geologically-directed, geochronological studies will address the degree to which these structural zones represent thrust planes and/or accretion boundaries. In addition, it is critical to establish if the regional structures are solely crustal features, or if any could represent fossil subduction zones.

If a structure taps the lower crust, or even the Conrad Discontinuity, it will serve as a suitable conduit along which CO₂-bearing fluids and some silicate magmas may pass from the mantle or lower crust into the upper crustal sequences. As discussed in Chapters 1 and 2 of Part I, regionally extensive zones of deformation are a characteristic of late tectonic development of granitoid - greenstone terranes and are the locus of gold mineralization and carbonate alteration. The extensive strike length of the Destor - Porcupine and Larder Lake fault zones in the Abitibi greenstone belt, from the Kapuskasing Structural Zone east to the Grenville front, implies that these two structures are profound crustal structures, which may have penetrated to at least the Conrad Discontinuity. Localization of fault-related carbonatization and gold deposits indicates that the regional zones of deformation played an important role in focusing the influx of CO₂-rich fluids into the upper crust (Chapter 2, Part I).

5.3.4. DISTRIBUTION AND VOLUME OF CO₂

The immense volume of CO₂ which was fixed in the southern part of the Abitibi belt (minimum of 490,900 moles of carbon; Fyon 1986) is present in regional and fault-related zones of carbonatization, localized by major zones of deformation. This volume and localization of carbonate implies the production of CO₂-bearing fluids was related to a process which operated at a scale much larger than that represented by the late intrusive centers. These factors suggest that sites of localized magmatism may be important controls on the development of gold mineralization on a camp scale (200 km²), whereas the extensive regional carbonatization must reflect a process which operated on a scale comparable to the exposed crust. As demonstrated in Chapters 2 and 3, Part I, the introduction of CO₂ and the emplacement of late plutons were coeval, and both events are manifestations of the process which triggered the cratonization of the crust. However, the conditions which led to the carbonate introduction chronology, observed in many late zones of deformation (Chapter 6, Part I) are not understood; they may provide a key to the compositional evolution of the fluid and hence have a significant bearing on understanding of the gold transport mechanism.

5.3.5. ASSOCIATED SILICA-UNDERSATURATED INTRUSIONS

Potassic and sodic plutonic complexes, showing varying degrees of silica-saturation, are often proximal or intimately associated with some gold deposits (Chapter 4, Part I) and were emplaced late in the plutonic history of greenstone belts (Chapters 4 and 5, Part I). The presence of late, silica-undersaturated igneous rocks can be ascribed to melting of anhydrous mantle peridotite in the presence of a CO₂-rich vapour (Mysen and Boettcher 1975). While a genetic link with gold may be speculative, on the basis of geochemical (Cameron and Carrigan 1987, Cameron and Hattori 1987, Rock *et al.* 1987) and geochronological data (Marmont and Corfu 1988), the presence of alkaline magmas may simply indicate that the mantle served as a source for some intrusions, but did not directly contribute to the gold metallogenic event.

5.3.6. SODIC AND POTASSIC ALTERATION

Sodium and potassium metasomatism of wall rock is a common characteristic of Archean gold mineralization (Chapter 3, Part I). Albite is a common accessory mineral in and immediately adjacent to many Archean, gold-quartz vein systems. Potassium feldspar is known to occur in wall rock alteration assemblages which have developed in gold deposits where felsic rock is abundant (eg. Hemlo; Kuhns 1986). Potassium-rich mica (sericite or biotite) is present in virtually all gold deposits. The introduction of Na and K into the gold systems may also be a consequence of mantle-derived, CO₂-catalyzed, lower crustal granulitization. Incongruent dissolution of feldspars, under conditions appropriate for granulite metamorphism, is strongly enhanced in the presence of a fluid having a high CO₂/H₂O ratio, with the ion solubilities being sodium > potassium >> calcium (Glassley 1983). The presence of CO₂-rich fluid in the lower crust should produce Ca- and Al-enriched, and Na-, K-, and Si-depleted residues during granulitization, consistent with the known geochemistry of granulite terranes (Chapter 4, Part II). Some, if not all, of the Na and K released during the granulitization of the lower crust may become incorporated into siliceous partial melts, generated at higher crustal levels.

5.4. GOLD SOURCE

A potentially very large gold reservoir is provided by that part of the lower crust undergoing granulitization. The only study to address the distribution of gold in granulites (Sighinolfi and Santos 1976) found that granulites from Bahia State, Brazil, contain 0.8 ppb gold (log transformed data excluding 4 "anomalous" data which exceeded 8 ppb). Inclusion of the four apparently anomalous samples (8.1, 16.0, 16.2, 18.4 ppb Au) only raises the log transformed mean to 0.9 ppb gold. These data indicate that the granulite rocks from the Bahia State are depleted in gold by a factor of at least 2 with respect to either fresh crustal igneous rocks or their lower metamorphic grade equivalent lithologies, an observation which cannot be made of any other metamorphic rock, on the same scale, using accurate analytical data (Crocket 1974, Paul *et al.* 1977). The 0.8 ppb gold abundance of these Brazilian granulites represents an upper limit, because 27 of the 101 samples analysed by Sighinolfi and Santos (1976) contained less than the detection limit of gold (0.4 ppb). These data suggest that gold can be removed from the lower crust during granulitization. Similarly, the incompatible element suite (eg. K, Rb, Li, B) which defines a dispersion halo about many Archean gold systems (Chapter 3, Part I) is identical to that which is depleted from many granulites (Chapter 4, Part II). Thus, gold and the companion incompatible element suite may have been mobilized from the lower crust, beneath parts of the Superior Province, during the late granulitization event, and concentrated in the supracrustal domains.

The distribution of gold in crustal rocks requires additional quantification, particularly as regards the transition up to and within granulite terranes. Such studies sometimes yield equivocal conclusions, primarily because of analytical error and poorly constrained sampling. However, the analysis of individual minerals or mobilizates, by high precision techniques, from well constrained samples, should provide clues to the behaviour and distribution of gold under conditions of granulite metamorphism.

5.5. GOLD TRANSPORT

At least two distinct pathways exist to extract and transport gold to the depositional site in the upper crust. Gold could have been partitioned into silicate magmas, generated by anatexis of either mantle or crustal material, and transported to the

upper crust where it, and companion elements, would partition into an exsolved, H_2O - CO_2 -bearing magmatic fluid. Alternatively, gold could have been transported by a vapour phase from the source area to the depositional site, without passing through the silicate magma medium. The essential difference between the two end-member models is the mode of transport. In the magmatic model, a silicate magma is the transport medium to the upper crust, whereas in the vapour phase model, the fluid is the transport medium. As emphasized in Chapter 3, Part II, the distinction between magmatic and metamorphic fluid in the lower crust is moot.

5.5.1. SILICATE MAGMA

Magmas generated within or deeper than the zone of granulitization provide optimal transport media, because they are generated within or must pass through the zone of gold depletion. Of the intrusions in the Superior Province that have been chemically characterized, only some pegmatites have compositional characteristics consistent with their derivation from the zone of granulitization (Figure 4.2, Part II). Tonalite-trondhjemite intrusions could have been produced above or below the zone of granulitization, and alkaline intrusions were derived from mantle sources (Chapter 4, Part II). Therefore, all intrusion types could satisfy the magma source requirements, but not necessarily by virtue of having been generated within the zone of granulitization.

On the basis of the temporal association between gold and silica-undersaturated intrusions (Marmont and Corfu 1988), and possibly anomalous gold concentration in some occurrences (Rock *et al.* 1987), the lamprophyre suite may be the preferred transport medium in a magmatic genetic model. However, alkaline magmatism may be just another manifestation of the overall cratonization process, emphasizing the involvement of mantle diapirism and CO_2 , and may not play a direct role in the transport of gold to the depositional site.

5.5.2. VAPOUR PHASE

Leakage of CO_2 -rich vapour from the zone of granulitization and diapirism into the upper crust can account for the introduction of large amount of CO_2 late in the magmato-thermal history of the crust, and for the gold - carbonate association. However, abundant CO_2 (manifest as carbonate) was introduced into the regional zones of deformation, before and during gold precipitation. This

carbonate introduction chronology (Chapter 6, Part I) may reflect changes in the redox conditions of the ambient fluid during granulitization, and/or complex fluid unmixing at deep crustal levels (Chapter 4, Part II). Changes in redox conditions may have transformed the ambient fluid into an oxidized, CO₂-rich, S-bearing (SO₂ ?) fluid, capable of scavenging and transporting gold from a large rock reservoir. "Leakage" of such auriferous, "magmato-metamorphic" fluids along diapiric pathways or deeply penetrating crustal structures into the upper crust provides a means of introducing CO₂ and gold into the supracrustal packages, later in the terminal, magmatic and thermal history of a greenstone belt, following the precipitation of the earlier, gold - poor carbonate.

5.5.3. DISTINCTION BETWEEN TRANSPORT MECHANISMS

Distinction between the magmatic and evolved vapour phase processes is not possible at present. Silicate magma transport of gold to upper crustal regimes may appear more appealing because of the temporal and spatial association between gold and late intrusions (Hodgson 1986), particularly those having an alkaline affinity (Marmont and Corfu 1988, Rock *et al.* 1987). This is analogous to some younger, Cordilleran-type, disseminated gold and copper-gold mineralization (Mutschler *et al.* 1985). Remaining uncertainties regarding the fluid include the evolution of the fluid present in the lower crust during and following peak granulite metamorphism. The contemporaneity between relatively oxidizing lower crustal hydrothermal activity, which post-dated peak granulite facies metamorphism, and hydrothermal activity within the auriferous systems, recorded in the hydrothermal mica dates in the upper crust (Chapter 4, Part II), suggests that a component of hydrothermal activity in the auriferous systems was derived from the lower crust, and accessed the upper crust during a period of apparent igneous quiescence. Late, post-peak metamorphic fluid activity in the lower crust is supported by titanite, monazite, and rutile age relationships (Corfu 1987, Corfu 1988, in press, Krogh *et al.* 1988). The influence of this fluid is unknown, but it was coeval with fluid activity in the auriferous zones (Figure 4.1, Part II). Thus, late fluid activity in auriferous zones may have been coupled with that in the lower crust.

While there are some constraints on the late thermal history of the lower crust, little is known of the character and origin of fluids

present during post-peak metamorphic period of cooling. To adequately assess possible fluid sources and describe fluid evolutionary histories, it is essential to characterize the late fluid activity recorded in the lower crust and in the auriferous zone near surface. Furthermore, it is necessary to document the magmatic fluids of different magmatic suites. At present, this aspect of Archean plutonism is essentially unknown. Only then can the fluid component of the gold metallogenic model be complete.

5.6. SUMMARY

The observed late magmatic and gold metallogenic evolution of blocks within the Superior Province can be modeled closely by a coupled sequence of lower crustal metamorphism and upper mantle partial melting, during which mantle-derived magma was emplaced at the base of and into the crust. The introduction of heat and CO₂-bearing vapour induced granulitization of the lower crust. Tonalite - trondhjemite magmas are produced by partial melting of existing lower crust (previously underplated basaltic magma?). Silica-saturated to -undersaturated magma was also directly produced by partial melting of the upper mantle; the melt composition having been influenced to some degree by CO₂ concentration in the ambient vapour.

Influx of CO₂-bearing vapour along major crustal structures into the upper crustal rocks culminated with intense carbonatization of susceptible lithologies and generation of nearly pure carbonate veins within regionally extensive deformation zones. Changes in the ambient redox conditions in the lower crust and upper mantle may have evolved to become oxidizing, and influenced the redox state of indigenous or transient CO₂-rich vapour or silicate magma. Oxidizing conditions may have enhanced the solubility of gold and sulphur. Incompatible elements (eg. LILE, U, Th, Zr, Au) may have been transferred from the zone of granulitization to the upper crust, either by vapour phase or silicate magma. Site-specific factors controlled gold precipitation in suitable, structurally induced, zones of permeability.

While unequivocal distinction between either the purely magmatic or purely vapour fluxing model cannot be made at present, it is important to emphasize that the gold metallogenic event is but one manifestation of a more profound magmatic and thermal history, which cratonized parts of the Superior Province. However, Archean gold metallogeny must be considered in the context of the totality of

PART II. GENETIC MODEL

the late Archean cratonization events. Appeal to any end member hypothesis may prove overly simplistic, as both processes played symbiotic roles in concentrating gold during the culmination of the cratonization event. Thus, it is most consistent to acknowledge that both magmatic and metamorphic process played a role in the concentration of gold, cognizant that the distinction between magmatic and metamorphic processes becomes less distinct at lower crustal depths. It is realistic to consider that a single source or even domain contributed all constituents to the hydrothermal system: radiogenic data illustrate that multiple sources contributed constituents to the hydrothermal system (eg. Kerrich *et al.* 1987). In the context of our model, multiple mantle and crustal sources are required. Thus, only by adopting a holistic perspective of the late magmatic, thermal and tectonic history of Archean shield areas are we likely to further our understanding of Archean gold metallogeny.

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